The Lateglacial history of Bassenthwaite Lake and adjacent areas, Lake District, UK; a sub-bottom profiling and geomorphological investigation

JENKINS, GERAINT,THOMAS-HOWARD

Geraint Jenkins

The Lateglacial history of Bassenthwaite Lake and adjacent areas, Lake District, UK; a sub-bottom profiling and geomorphological investigation

---

Geraint Jenkins

The Lateglacial history of Bassenthwaite Lake and adjacent areas, Lake District, UK; a sub-bottom profiling and geomorphological investigation

The glacial history of Bassenthwaite Lake and adjacent areas is poorly understood despite the English Lake District being a key area for assessing the dynamics of the British-Irish ice sheet. This project undertook geomorphological mapping and sub-surface geophysical investigations in order to investigate the glacial history of Bassenthwaite Lake. Drumlins and moraine ridges adjacent to the lake point to a relatively simple ice advance and retreat during the Last Glacial Maximum (LGM), but sub-bottom profile data from the lake floor suggests a more complex glacial history with a lower basement till or bedrock overlain by glaciolacustrine and glaciofluvial sediments. An eroded lower till or bedrock illustrates erosion potentially taking place during LGM ice coverage with the Bassenthwaite basin entirely inundated at this period. Moraine ridges along the length of Bassenthwaite Lake illustrate punctuated retreat of a constrained Lateglacial valley glacier. In addition to this, ice-marginal fluctuations are recorded through localised deformation within glaciolacustrine sediments. Continued retreat of the glacier results in the deposition of an upper glaciolacustrine sequence with ice-berg rafted debris identified. The identification of a re-advance till unit and the deformation of a lower glaciolacustrine sequence illustrates a fluctuating ice margin, which in some places re-advanced up to ~ 1 km and points to a dynamic ice mass occupying the Bassenthwaite basin. A drumlin was also investigated through sub-bottom profiling and was interpreted to have been created through erosion of the underlying bedrock. Following this, sedimentary packages overlie and infill the drumlin with two glacier re-advance events resulting in drumlinisation. It is clear that sub-bottom profiling has the potential to illustrate the internal composition and structure of submerged drumlins in addition to providing valuable information into Lateglacial glacier retreat dynamics.

The Lateglacial history of Bassenthwaite Lake and adjacent areas, Lake District, UK; a sub-bottom profiling and geomorphological investigation

Geraint Thomas-Howard Jenkins

Thesis submitted for the degree of Master of Science

Department of Geography

Durham University

March, 2015

Table of Contents

List of Figures

Statement of Copyright

Acknowledgements

Firstly, I would like to thank my supervisors Professor Chris Stokes and Dr Dave Roberts, who have been supportive, understanding and accommodating throughout the undertaking of my Masters degree. I could not have asked for better supervisors.

I would also like to thank Professor John Woodward of Northumbria University for the use of his equipment, explanation of its application and his presence during fieldwork. The collection of data from within Bassenthwaite Lake was also the result of hard work by Frank Davies and his technical team at Durham University.

Outside of the University department I would like to thank Jenny Geary and Wendy and Ian Jenkins, who always attempt to keep me afloat and sailing in the right direction.

1. INTRODUCTION

Using a combination of geophysical and geomorphological investigations, this study aims to explore and reconstruct ice sheet and glacier retreat dynamics during the LGM in the vicinity of Bassenthwaite Lake in Cumbria. This area is particularly interesting because i) the glacial history of the area is poorly explored, and ii) a series of drumlins adjacent to and potentially within the lake provide an opportunity to reconstruct subglacial conditions during the last glacial cycle.

Understanding the extent and behaviour of ice sheets and glaciers is important to guide future predictions of ice retreat and therefore potential contributions to sea-level change (Pritchard *et al*., 2009). Ice sheets and glaciers contain ~ 70 % of the Earth's entire freshwater. An input of such large quantities of freshwater into oceans, due to changes in ice mass dynamics, has been illustrated to have a profound impact upon the global climate and ocean circulation systems (Rahmstorf, 2002). Further to this, a continuation of current ice mass loss has been projected to potentially result in a 0.5 to 1.4 metre contribution to global sea-level rise by 2100 (Rahmstorf, 2007). The need to investigate the dynamics and processes of ice sheets and glaciers is therefore evidently clear.

Ice sheet and glacier behaviour is influenced by conditions in the subglacial environment (Paterson, 2001), which are difficult to observe, measure and investigate beneath contemporary ice masses. To circumvent the problem of limited access to current glaciers, it is possible to examine landforms which are now exposed in formerly glaciated environments and this approach has been applied in numerous investigations (Clark *et al*., 2003; O'Cofaigh *et al*., 2005a). Understanding how ancient ice sheets behaved in relation to changes in climate can allow a greater understanding of contemporary ice mass-climate interactions. Further to this, the distribution and morphology of features which remain following ancient ice sheet recession can allow an understanding of potential ice processes occurring in current ice sheets. This aids interpretations of future ice-ocean-climate interactions. Landform assemblages and sedimentary evidence exists on ancient ice sheet beds and these can be used to infer palaeo-ice mass processes.

Landforms and groups of landforms (landform assemblages) have been shown to illustrate the extent of palaeo-ice sheets, occurrence of meltwater and ice stream flow phasing (Stokes and Clark, 2001). They have also been considered to have the potential to illustrate palaeo-ice velocity (Stokes and Clark, 2002). Exposures within palaeo-ice sheet beds also have the potential to provide vital process information through the composition and structure of sediments. Sediments have the potential to record numerous palaeo-ice mass processes such as former ice extents, the presence of meltwater, and the potential for subglacial ice flow. Therefore, sediments and landforms provide vital information regarding subglacial processes and palaeo-ice mass dynamics which can aid in the consideration of contemporary ice sheet processes.

The investigation of subglacial landforms allow us to improve our understanding of flow mechanisms and processes of erosion and deposition. However, in some locations, we tend to only have limited exposures and some landforms can also be submerged. Investigations using geophysical techniques (Burke *et al*., 2008; Hiemstra *et al*., 2011), however, have demonstrated the potential for examining the sub-surface composition and structures of landforms on palaeo-ice mass beds. The application of geophysics allows the consideration of large-scale, 3-dimensional sedimentary composition and structure. This is particularly important for subglacial bedforms because small exposures may not illustrate the entire, full composition and structure of the feature. One such area of submerged, glacial landforms is the English Lake District, which is also a key area for the dynamics of the British-Irish ice sheet.

This thesis presents new data on the geomorphology and sedimentology from one of the largest lakes within the Lake District. This study aims to provide a proof-of-concept investigation for future applications to drumlins in addition to investigating the Lateglacial history of Bassenthwaite Lake. Furthermore, no geophysical or sedimentological investigations currently exist for Bassenthwaite Lake despite this location containing vital information in regard to glacier retreat dynamics. Aside from an investigation of glacier retreat dynamics by Pinson *et al*. (2013) for Lake Windermere, very few sedimentary and geomorphic investigations have been undertaken within the Lake District. Furthermore, lakes such as Lake Windermere have shown that they contain glacial sedimentology which may not be recorded outside of the lake environment. Large water bodies also submerge vast swathes of glacier retreat locations and may therefore contain vast information regarding potential subglacial processes and valley glacier dynamics.

1.1. Aims & Objectives

The aim of this project is to investigate the glacial history of Bassenthwaite Lake and adjacent areas, using geophysical analysis and the identification of the glacial geomorphology. In addition to this, the sub-surface composition and structure of drumlins may be obtained. The combination of internal sedimentology and geomorphology will aid interpretations of potential palaeo-ice mass processes and the glacial landsystems of Bassenthwaite Lake. The three principle objectives of this study are to:

1. Use sub-bottom profiling images to identify and investigate the topography of the bed of Bassenthwaite Lake to help identify glacial landforms
2. Use sub-bottom profiling images to observe the sedimentary architecture of the landforms identified from the lake bed topography

3. Use both (1) and (2) above to help reconstruct the glacial history of Bassenthwaite Lake in conjunction with geomorphological mapping in the vicinity of the lake.

2. A REVIEW OF SUBGLACIAL PROCESSES AND LANDFORMS AND THE APPLICATION OF GEOPHYSICS

2.1. Ice sheets and their importance in the global climate system

The Greenland and Antarctic ice sheets hold a sea-level equivalent of ~ 70 metres and there are concerns about recent accelerations in mass loss as a result of climate warming (Alley *et al*., 2005; Shepherd *et al*., 2012; Bamber and Aspinall, 2013). Ice streams (corridors of faster flowing ice within ice sheets) play an important role as drainage mechanisms for ice sheets. This is illustrated by Antarctic ice streams, which discharge ~ 90 % of all ice and sediment from the Antarctic Ice Sheet (Bennett, 2003). Therefore, the processes involved with these areas of fast ice flow need consideration when attempting to understand ice sheet dynamics. The majority of movement within ice streams and glaciers occurs through subglacial motion and, unfortunately, these areas are difficult to observe at modern ice streams. Investigations instead are concentrated upon ancient ice sheet beds, with sediments and landforms in these locations providing potential information on ancient subglacial processes.

2.2. Palaeo-ice sheets and subglacial processes

Sediments and landforms deposited by ancient ice sheets provide the potential to understand former ice sheet processes (Bennett and Glasser, 2011). This can then allow a consideration of ancient ice sheet response to changes in climate. This may then allow a consideration of how current ice sheets may respond to climatic alterations. Furthermore, the geomorphic-sedimentary record provides an excellent opportunity to study subglacial bedforms and therefore allows interpretations of subglacial processes (Clarke, 2005). Current investigations of subglacial processes, however, can be supplemented by studies on palaeo-ice sheet beds. Therefore, sediments and landforms deposited across the UK provide vital information regarding subglacial and deglacial processes for the British-Irish Ice Sheet that are likely to help inform our understanding of modern ice sheet dynamics.

Ice sheet behaviour is heavily influenced by conditions at the subglacial environment (Paterson, 2001; Clarke, 2005). The processes operating beneath glaciers are considered to have a larger influence on overall ice dynamics than those operating within them (Clarke, 2005). Subglacial processes, however, are difficult to observe, measure and investigate beneath current ice masses.

Thermal conditions at the ice-bed interface and the physical properties of the underlying substrate influence which subglacial processes are likely to operate (Clarke, 2005; Evans *et al*., 2006). Optimum conditions for subglacial movement are considered to be warm-based ice, over a soft substratum, allowing deformation of the bed (Clarke, 2005). However, rapid basal flow is also recognised over hard bedrock types if there is abundant, high pressure, subglacial meltwater (Clarke, 2005). High pressure meltwater is also important for the mobilisation and transport of subglacial sediment (Boulton, 1996; Clarke, 2005). These 'soft' sedimentary glacier beds and how subglacial water may interact with them is an important component that is difficult to capture in numerical models of ice dynamics (Smith *et al*., 2007). Unfortunately, observations of these processes are scarce and models may therefore fail to illustrate actual ice mass behaviour. Recent observations in West Antarctica suggest that subglacial processes can be spatially highly variable (Smith *et al*., 2007). Furthermore, not all processes below ice masses are fully understood (Clarke, 2005). These include the depth to which deformation may take place within a subglacial sediment if present (Truffer *et al*., 2001) and the efficiency of subglacial drainage (Gray *et al*., 2005). These further stimulate the need to investigate former ice sheet beds and interpret former subglacial processes from the geomorphic-sedimentary record. Indeed, it is well known that such processes leave a suite of landforms beneath ice sheets, collectively referred to as subglacial bedforms.

2.3. Subglacial bedforms

Subglacial bedforms are created and modified below ice sheets and glaciers and can be longitudinal, transverse and even hummocky in morphology (Eyles *et al*., 1999; Benn and Evans, 2010). It is common practice to use geomorphological features to reconstruct former ice dynamics and processes (Clark, 1997). Subglacial bedforms are particularly important as they can provide information regarding flow patterns, thermal regime, subglacial processes and ice velocity (Hart, 1999; Stokes and Clark, 2002; O'Cofaigh and Stokes, 2008). It is also documented that certain subglacial bedforms can be associated with specific subglacial conditions (Kleman and Hattestrand, 1999). Therefore, understanding their spatial distribution may provide insights into former glacier behaviour and in particular thermal regimes.

Longitudinal streamlined bedforms, aligned parallel to ice flow can be categorised into flutings, drumlins and mega-scale glacial lineations (MSGLs). Such categories are based on the elongation ratio and length of the features with the elongation given as:

$$E = l_b / w_b$$

where, l_b is the maximum subglacial bedform length and w_b is the bedform maximum width. Such longitudinal features, however, have been considered a continuum (Rose, 1989) and definitions between bedforms are perhaps drawn arbitrarily. Drumlins, in particular, have been extensively studied regarding the potential for the extrapolation of subglacial processes.

Drumlins are considered to be the most extensively investigated subglacial feature (Menzies, 1979; Stokes *et al*., 2011). Drumlins are defined as a streamlined, oval feature with the long axis parallel to the orientation of ice flow (Stokes *et al*., 2011). Not all drumlins are observed to be oval in morphology however, and a continuum of drumlinoid forms has been described (Benn and Evans, 2010). These include wider, asymmetrical drumlins (parabolic forms) (Spagnolo 2010; 2011; 2012) and longer, narrow drumlins (spindle forms) (Benn and Evans, 2010). Complex drumlinoid bedforms also exist which include 'transverse asymmetrical drumlins' and 'superimposed drumlins' which occur deposited over younger bedforms (Rose, 1989). The importance of drumlins are detailed by Baranowski (1979: p. 435) who writes: "until the mechanisms responsible for drumlin formation are fully understood, some of the key glaciological problems related to the glacier bed will remain obscure". Recent geophysical investigations beneath the Antarctic ice sheet have confirmed the formation of drumlins beneath the ice (King *et al*., 2007). However, drumlin formation remains controversial with numerous hypotheses proposed (Smalley and Unwin, 1968; Shaw *et al*., 1989; Hindmarsh, 1998; Fowler, 2000; Schoof, 2007; Stokes *et al*., 2011). Such varied hypotheses result, in part, from the spatial inconsistency of internal structure and composition (Patterson and Hooke, 1995).

Drumlin internal composition (constituents within the feature) and structure (arrangement of these constituents) can allow interpretations of formation conditions. This can aid in understanding subglacial processes. Drumlin internal composition and structure are variable, but most are observed to consist of till (Stokes *et al*., 2011). Some have been observed to contain rock cores which have subsequently been covered in a superficial carapace of till (Hart, 1995). A recent review by Stokes *et al*. (2011) outlined that drumlins can be classified into five commonly occurring compositions. These are: (1) mainly bedrock, (2) part bedrock / part till, (3) mainly till, (4) part till / part sorted sediments, and (5) mainly sorted sediments (Stokes *et al*., 2011) (Figure 2.1).

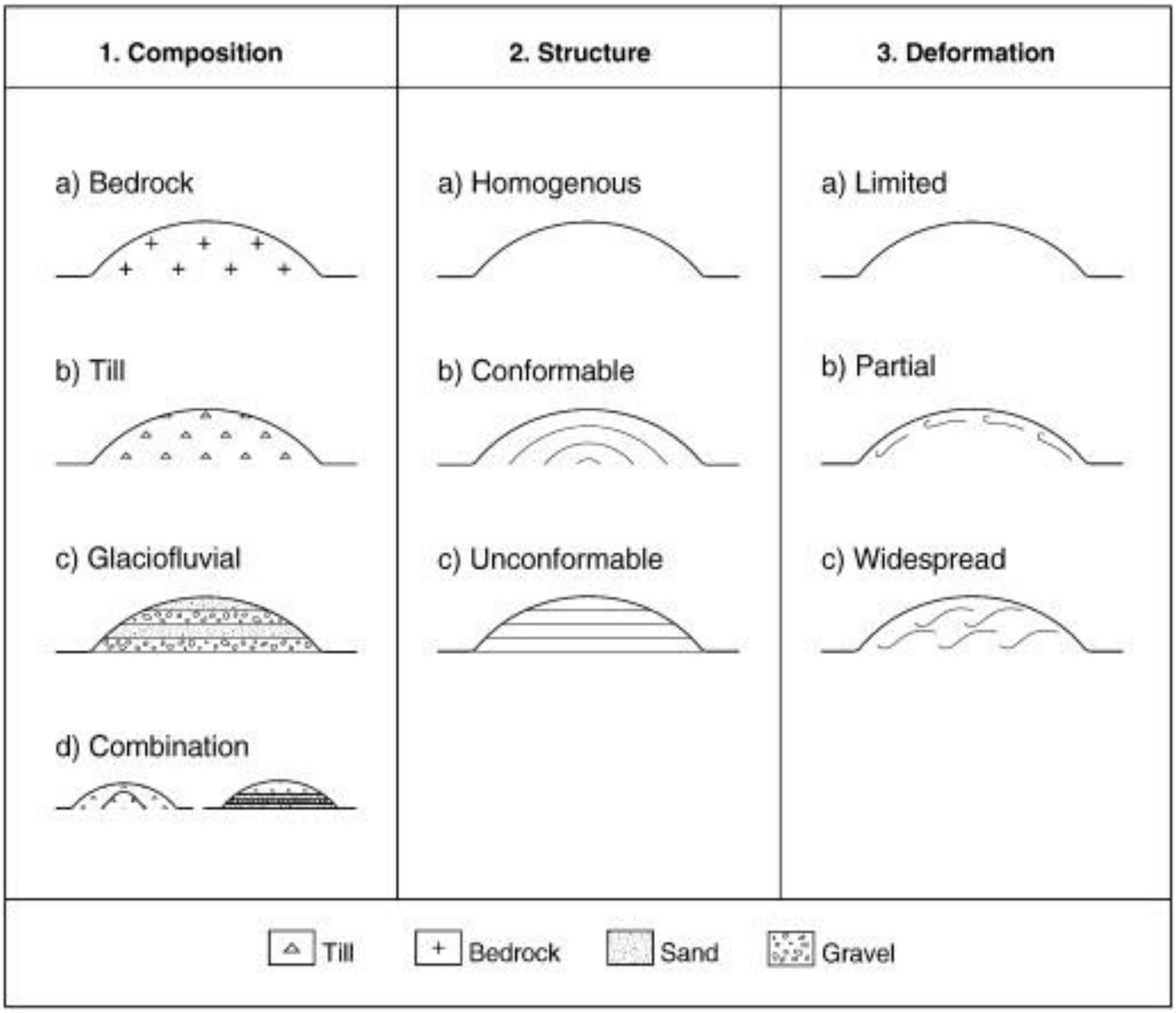


Figure 2.1. An example of the variation in composition, structure and deformation within drumlins (Stokes *et al*., 2013). Such differences might explain why there are a large number of formational mechanisms hypothesised in the literature.

Numerous investigations have identified bedrock or rock drumlins. These are defined as streamlined bedforms with an *entire* composition of bedrock (Heroy and Anderson, 2005; Kerr and Eyles, 2007). Rock drumlins have been observed to occur in various lithologies (Graham *et al*., 2009). Rock drumlins differ from other subglacial features such as roche moutonnée (because they are not plucked on the lee side) and whalebacks (as whalebacks are typically symmetrical in the long profile and drumlins are generally not). A number of drumlins with an internal composition of both bedrock and till have been recorded (Hill, 1971; Fuller and Murray, 2002). A number of these bedrock drumlins have been recorded with an overlying drape of till (Fisher and Spooner, 1994; Nenonen, 1994). Most often, rock cores within drumlins are observed near the stoss end and are termed 'rock-cored' drumlins (Boyce and Eyles, 1991; Fuller and Murray, 2002). The internal association of bedrock and till has been shown to be variable. However, it has been observed that a rock core surrounded by a single unit of till most commonly occurs (Meehan *et al*., 1997). The 'rock core' within drumlins has also been shown to shear into overlying till units (Meehan *et al*., 1997). Numerous other 'till-bedrock' associations exist and include drumlins consisting of multiple till units in addition to bedrock material (Fisher and Spooner, 1994).

Till composed drumlins have been shown to vary from a single unit of till within the bedform to a feature with clearly conformable layers (Hart, 1997; Menzies *et al*., 1997; Nenonen, 2001; Johnson *et al*., 2010; Figure 2.2; Figure 2.3). The composition and structure of till composed drumlins is highly variable. These include a single till unit with widespread deformation, two till units or more (Figure 2.4, Fuller and Murray, 2002) and two till units of varying ages (Rattas and Piotrowski, 2003). A key issue regarding the interpretation of till-filled drumlins is their potential conformability with the underlying surface. A conformable contact to the underlying geology potentially illustrates incrementally deposited sediment. An unconformable contact could therefore allow interpretations of erosional formation conditions (Stokes *et al*., 2011).

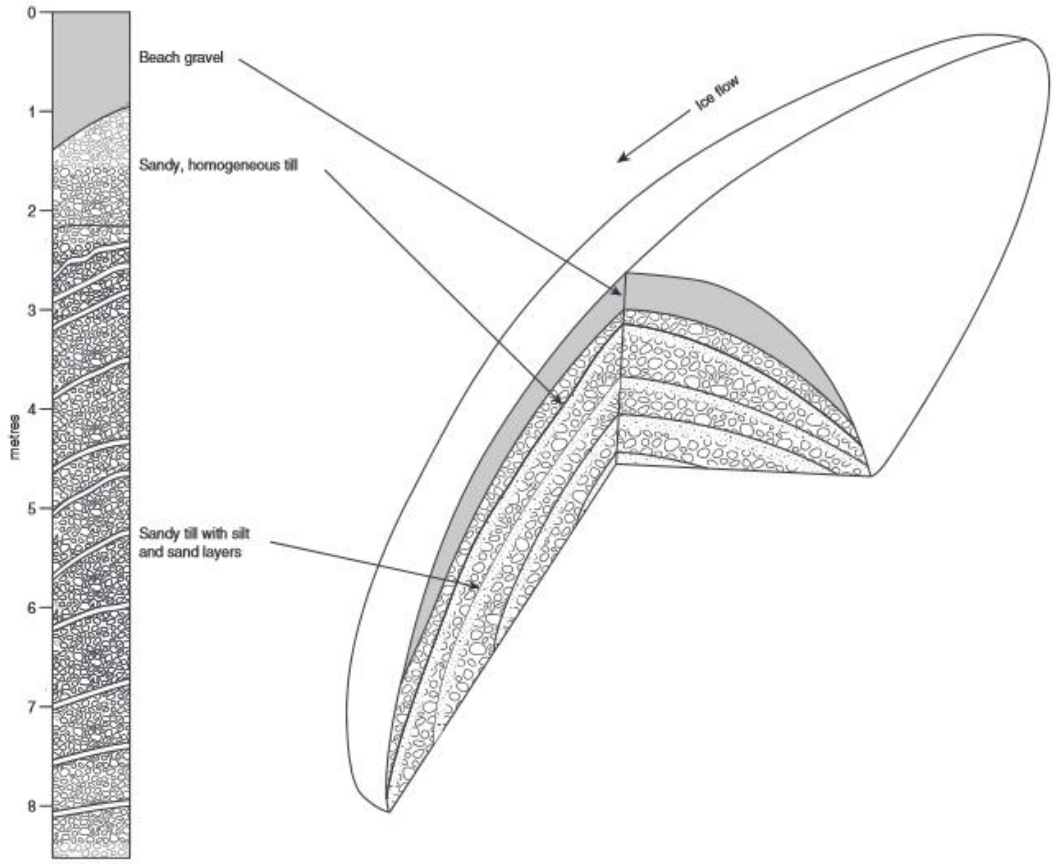


Figure 2.2. Sedimentary architecture of the Kaituri drumlin, illustrating an entirely till composed feature (Nenonen, 1994).

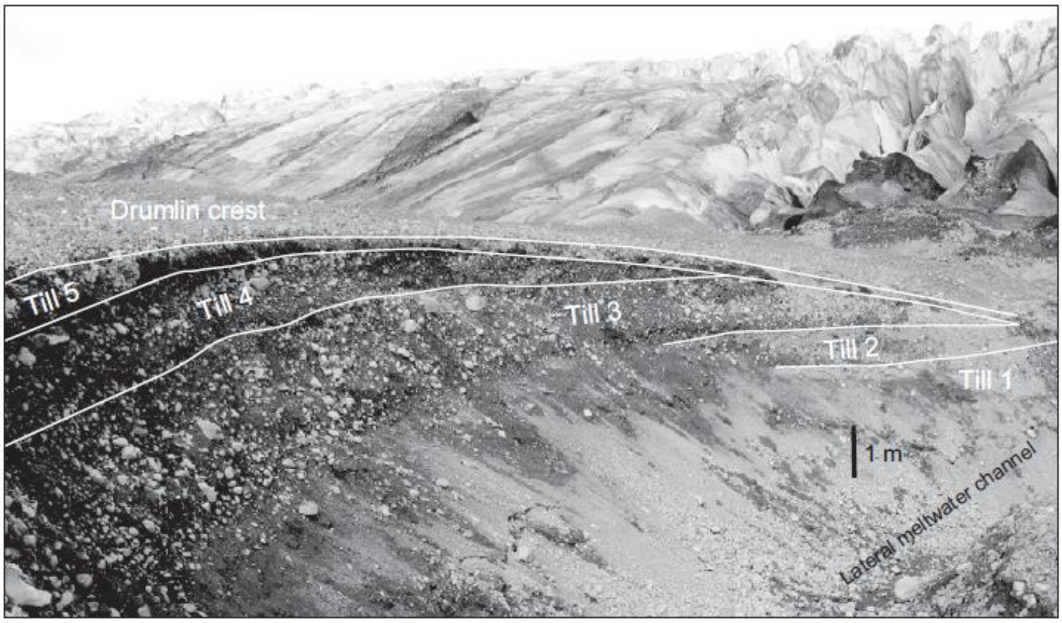


Figure 2.3. Image of five till units recorded in an active drumlin field revealed at the margin of the surge-type glacier, Mulajokull, Iceland (Johnson *et al*., 2010). This includes an overlying, drape till lithofacies followed by 4, ~ 1 m thick lithofacies which are exposed by a meltwater channel emanating from the snout of the glacier.

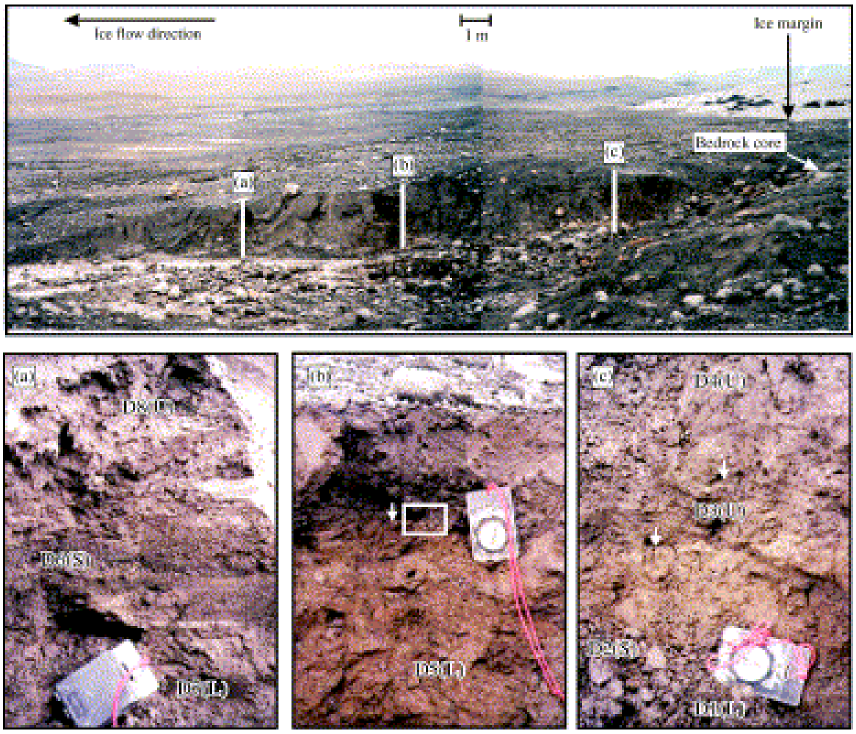


Figure 2.4. Photograph of a stream-cut drumlin exposure, Iceland (top image) with 3 stratigraphic logs (a – c) (Fuller and Murray, 2002). An upper and lower till were identified with clay bands and sand lenses also present.

Another frequently reported internal composition are drumlins containing both till and sorted sediments (Rattas and Piotrowski, 2003; Raunholm *et al*., 2003; Kerr and Eyles, 2007; Hiemstra *et al*., 2011). These include the sorted sediment acting as a pod or core, or the sorted sediment underlying till units (Clapperton, 1989; Jorgensen and Piotrowski, 2003). Sorted sediment has also been observed between till units (Kerr and Eyles, 2007) and inter-bedded with till (Goldstein, 1994). Drumlins composed of entirely sorted sediments have also been documented. These include a drumlin reported by Menzies and Brand (2007) that was exposed along its long-axis. Entirely sorted sediment compositions have led to interpretations of an infilling of sediment into subglacial cavities resulting from the presence of large quantities of meltwater (Stokes *et al*., 2011).

Observations of the entire composition and structure of a drumlin are rare, largely owing to a lack of large lateral exposures and geophysical investigations. Such an exposure of a drumlin is, however, reported by Menzies and Brand (2007) in the New York drumlin field, USA (Figure 2.5). This exposure allowed an interpretation of the conditions that pre-dated the drumlin and were interpreted to reflect sedimentary deposition in a deltaic environment. This was then followed by overriding, advancing ice, resulting in 'drumlinisation' (Figure 2.6). Such examples, however, are rare and further illustrate the need for geophysical investigation to provide large-scale internal composition data-sets. Geophysical analysis circumvents the problem of limited exposures within a drumlin. Such limited access to the internal composition of drumlins can result in assumptions of entire bedform composition and structure without direct analysis.

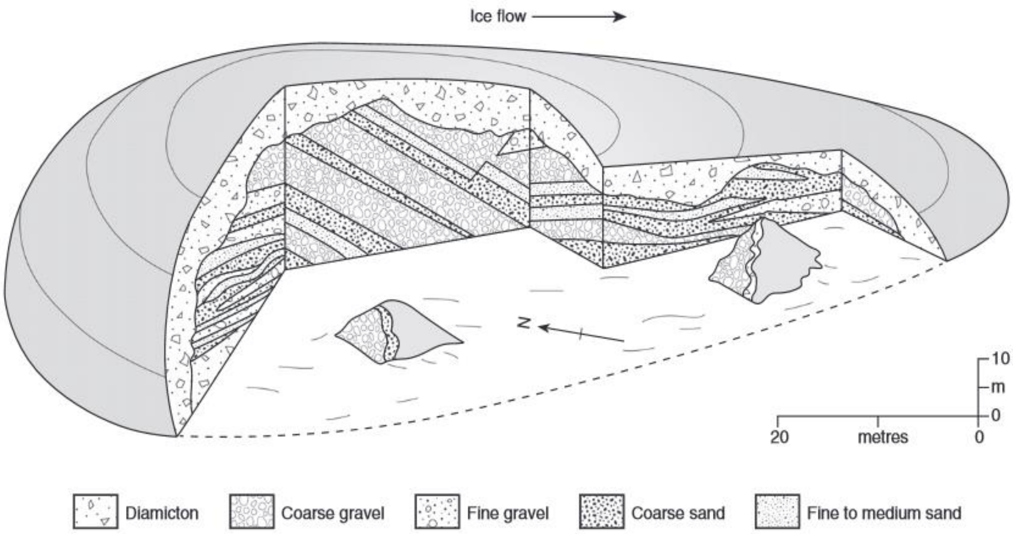


Figure 2.5. Internal composition and structure for a drumlin on the New York drumlin field, USA. Here, the drumlin is composed of mostly sorted sediments and allows, owing to its large lateral exposure, an interpretation of the entire feature. This allows more accurate interpretations of potential ice sheet processes (Menzies and Brand, 2007).

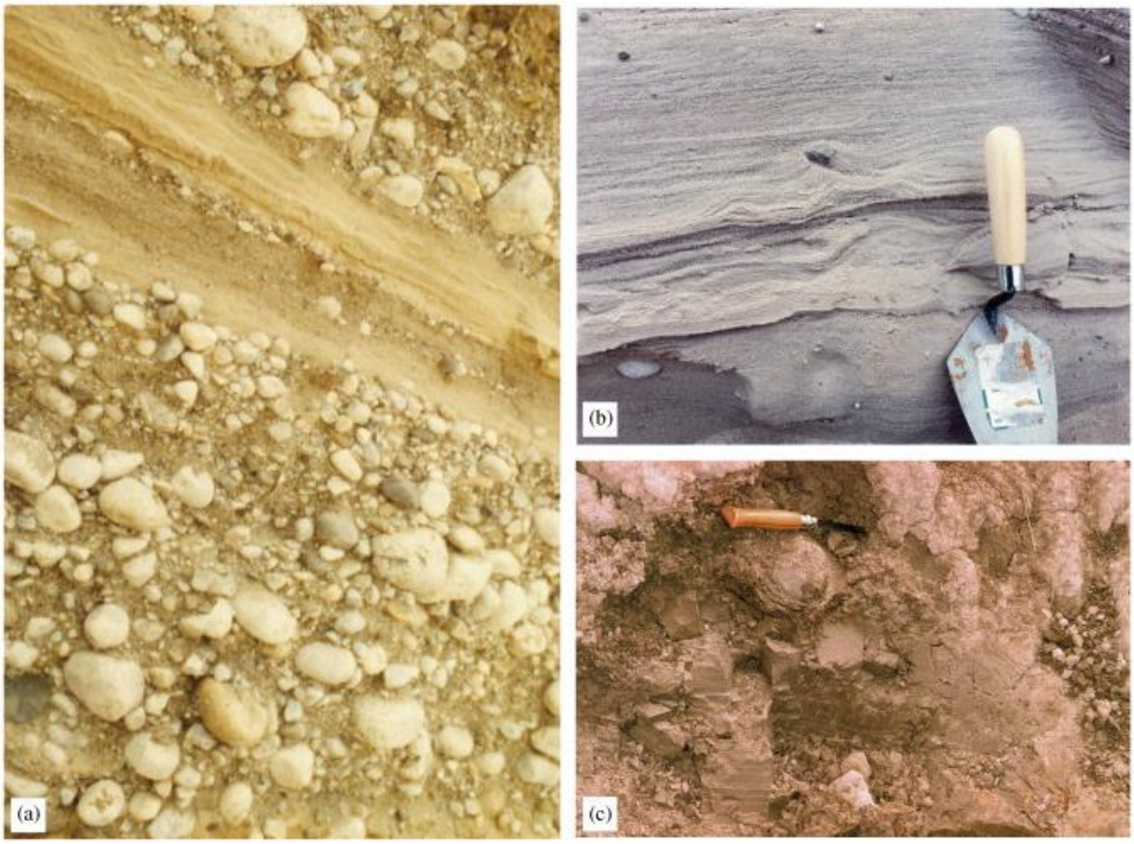


Figure 2.6. Steeply dipping, interbedded sand and gravel (a), medium to fine sands (b) and a reddish brown till reported within the drumlin illustrated in Figure 2.5 (Menzies and Brand, 2007).

The large spatial variation in drumlin internal composition and structure has potentially hindered a unifying theory for their creation (Benn and Evans, 2010; Stokes *et al*., 2011). Sugden and John (1976 p.239) state that “there are almost as many theories of drumlin formation as there are drumlins”. Internal drumlin composition and structure have been observed to vary between individual drumlins within the same drumlin field. This has led to the interpretation that drumlins may be formed by one of the following: (1) different drumlins are created by different subglacial processes despite their similar morphology, or (2) the same subglacial processes occur across large areas to produce a terrain that is drumlinised but with drumlins varying in internal composition and structure (Stokes *et al*., 2011).

A hypothesis for drumlin formation was proposed by Boulton (1987) and was based on the principles of preferential sediment erosion and distribution within a subglacially deforming layer. Here, the stronger areas of sediment (owing to larger grain size and therefore reduced pore-water pressure and high sediment strength) would result in ‘sticky spots’. The subsequent streamlining of these sticky spots would result in drumlinisation (Figure 2.7). This hypothesis also explains the presence of a ‘core’ within drumlins. These cores form resistant regions to which further sedimentation would be concentrated around (Benn and Evans, 2010).

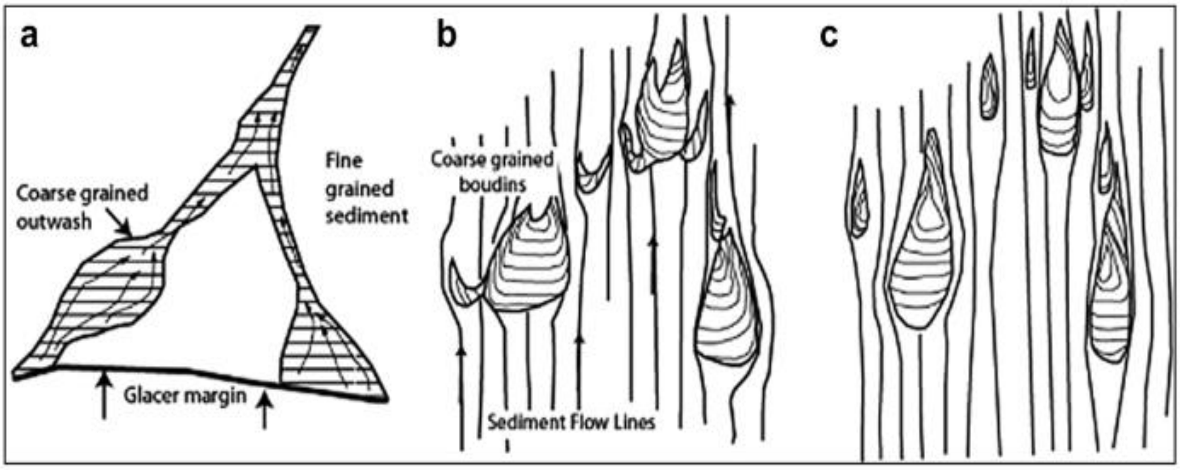


Figure 2.7. Illustration of progressive drumlin formation resulting from differentiations in the rheological properties of sediment (Boulton, 1987).

Kjaer *et al*. (2003) illustrates a similar mechanism for drumlin formation from the proglacial area of Slettjokull, Iceland. Here, it is shown that the thickness of a stratigraphically distinct till sheet is correlated to the substratum it overrides. A thicker till deposit occurs over a till substratum compared to a glaciofluvial, sorted sediment substratum (Benn and Evans, 2010). This results in the spatial distribution of till deposition influenced by the glacier forefield conditions. This repeated ‘till over till’ deposition is interpreted to result in the formation of proglacial drumlin-shaped features at the Slettjokull glacier forefield, Iceland (Figure 2.8).

Due to the presence of sorted and stratified sediments *without* evidence of deformation, some hypotheses suggest drumlin formation by sedimentation in subglacial cavities (Shaw and Kvill, 1984; McCabe and Dardis, 1989; Dardis and Hanvey, 1994). Cavity infill processes were proposed for a late Pleistocene drumlin in northwest Ireland by Dardis and Hanvey (1994). A suite of processes were identified from the sedimentary record. High energy, closed conduit systems where represented by pebbly sand facies. Closed conduits with a lower energy due to decreased meltwater were represented by sand and silt lithofacies and finally, stratified diamicts where interpreted to result from deposition along subglacial cavity margins (Dardis and Hanvey, 1994). Such variations in meltwater flux were also considered to result in a decoupling at the ice-bed interface. A complex system of erosion, deposition and meltwater flux was therefore interpreted to result in drumlin formation. A similar interpretation of subglacial processes and drumlin formation is provided by Knight (2014). Here, sedimentation is interpreted to have occurred in leeside, subglacial cavities for a drumlin in Connemara, Ireland (Figure 2.9). In this interpretation, the subglacial hydrological system and sedimentary processes are considered of most importance for drumlin formation.

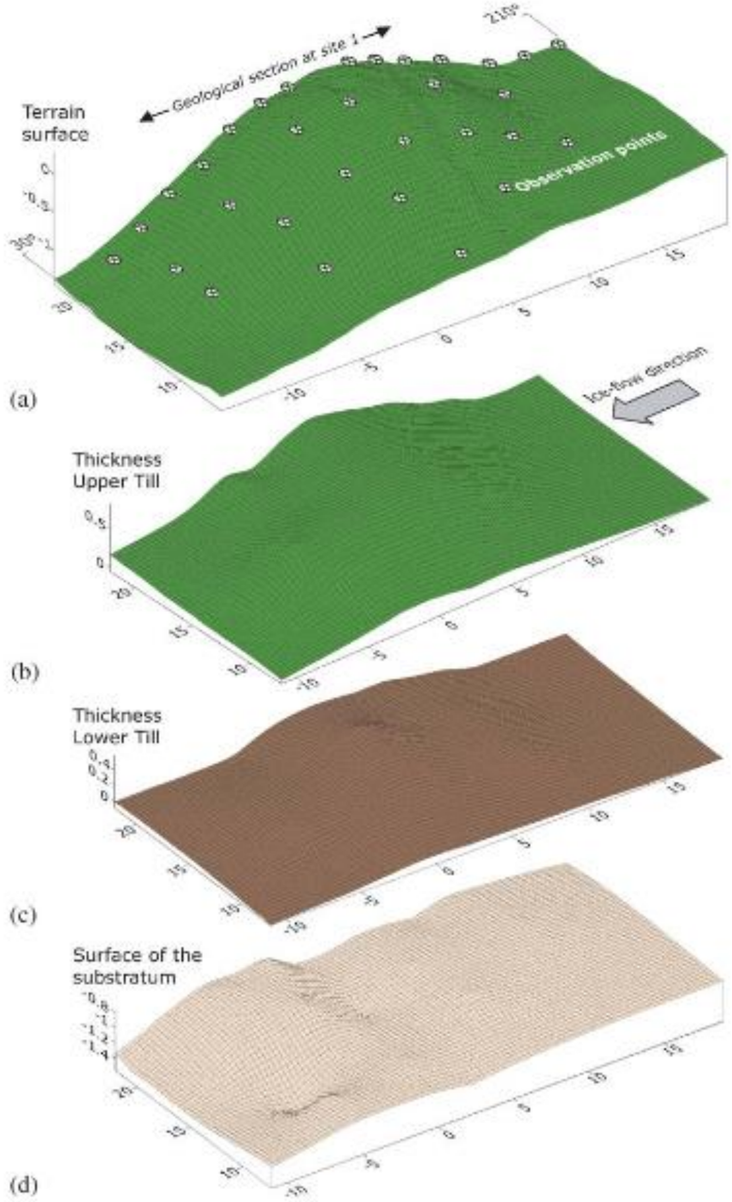


Figure 2.8. Internal sedimentary architecture of a potential drumlin feature at the forefield of Slettjokull glacier, Iceland. Here, the original relief of the substratum is given (d), with subsequent deposition of different till units (c & d) resulting in the present geomorphology of the drumlin feature (a) (Kjaer *et al*., 2003).

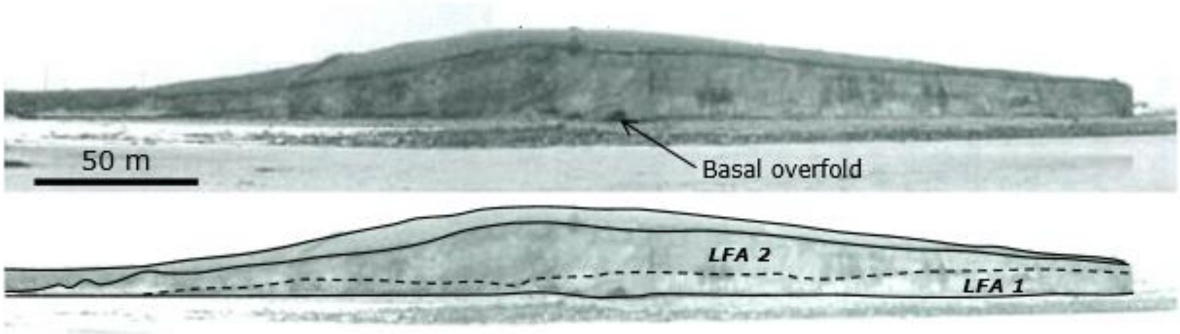


Figure 2.9. Panoramic image and sedimentary architecture interpretation of Ballyconneely drumlin, Ireland (Knight, 2014). Lithofacies association 1 is interpreted as a lodgement diamicton with lithofacies association 2 interpreted as a stratified debris flow deposited into a lee side, subglacial cavity.

Another potential hypothesis for drumlin formation has been proposed by Shaw and colleagues (e.g. Shaw *et al*., 1984; Shaw and Gilbert, 1990). It is suggested, in these studies, that catastrophic subglacial meltwater floods provided the conditions to create drumlins. This is interpreted from the composition of drumlins containing pre-existing sediment or bedrock which has subsequently been eroded into. A problem of this hypothesis, however, is that drumlin creation would require a large subglacial meltwater input. This would mean that large subglacial lakes would have existed across the bed of the ancient ice sheets. These water bodies would have influenced overall ice sheet behaviour and dynamics, but there is a dearth of sedimentological records for their occurrence. Thus, opposition to this hypothesis of drumlin formation has been argued (Clarke *et al*., 2005) primarily due to the lack of sedimentological and geomorphological evidence for vast, subglacial water bodies.

In contrast to sediment deposition and meltwater hypotheses, recent hypotheses have considered drumlin formation through numerical models. These generate an undulating surface from the interaction between the ice and a deformable bed. It is interpreted from model results that bedforms can be generated without alterations in sediment properties (Hindmarsh, 1998). Alterations in till thickness and therefore changes in effective pressures can lead to minor perturbations within the till and therefore the potential growth of obstacles and bumps at the subglacial interface (Hindmarsh, 1998). This ‘instability’ model has been developed further (Fowler, 2000; Schoof, 2007) and describes how waves (drumlins) can be produced at the ice-bed interface (Figure 2.10; Stokes *et al*., 2013). Here, it is proposed that the coupled flow of both ice and subglacial till results in spontaneous landform creation through relief change in the subglacial till. An ‘unstable’ system is therefore invoked whereby positive feedbacks result in the amplification of small disturbances to create larger forms (Stokes *et al*., 2013).

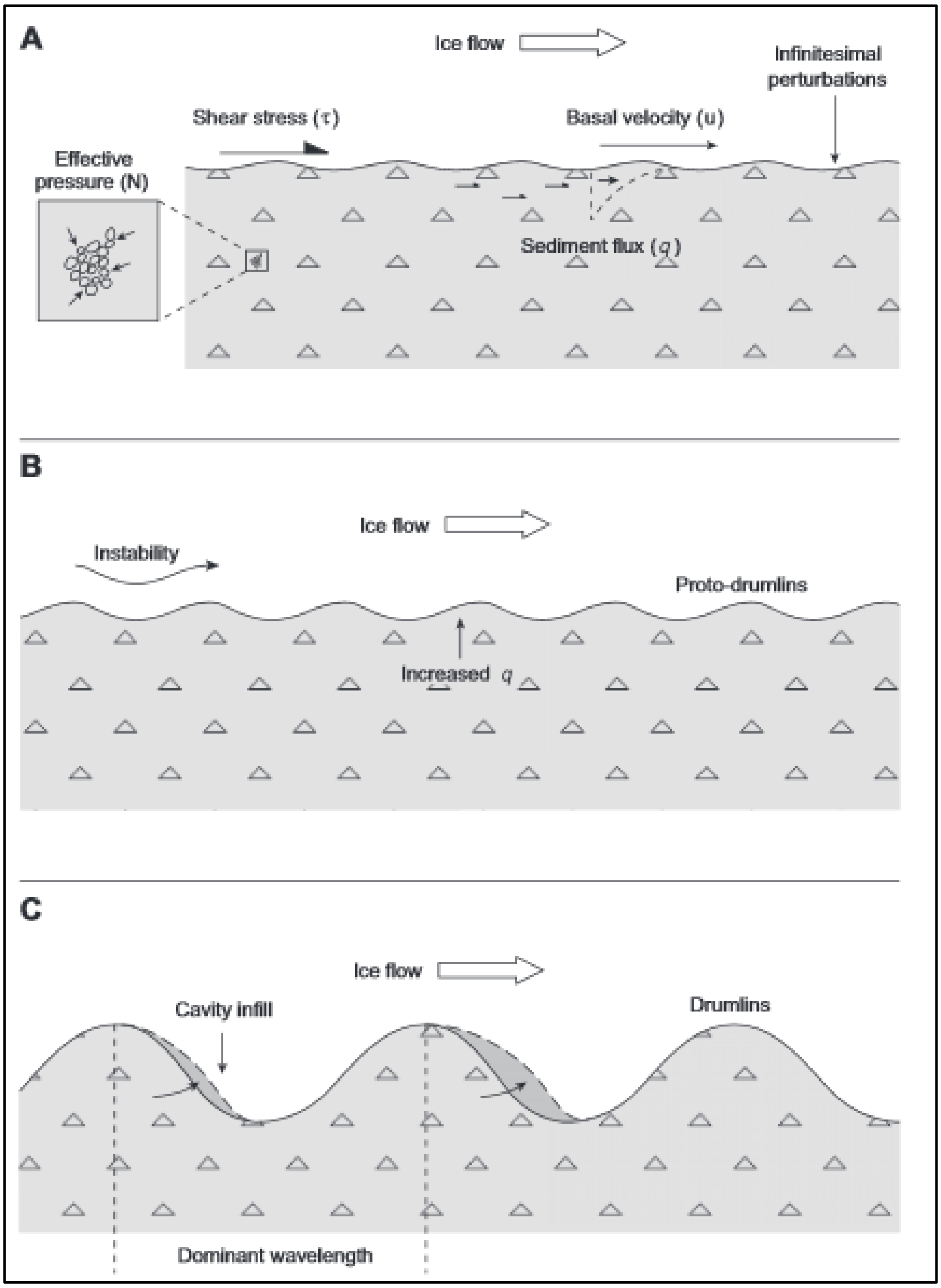


Figure 2.10. Temporal illustration (a - c) of the proposed instability hypothesis for drumlin formation. Ice and sediment deform simultaneously and develop along-flow instabilities, resulting in waveforms (Stokes *et al*., 2013).

2.4. Application of geophysics

Geophysical analysis of the evolution of bedforms has been undertaken on modern ice streams and allowed interpretations of potential processes (King *et al*., 2007; Smith *et al*., 2007). This provides advantages over the investigation of landforms created by ancient ice sheets as both timescales and methods of creation can be observed in 'real-time'. However, such investigations are in their infancy because of the logistical difficulties and the expense involved with geophysical analysis. Problems, however, exist in this type of investigation due to the need to penetrate relatively thick ice masses and undertake repeat investigations over numerous field-sessions. Further to this, to understand entire ice stream-landform processes, a large-scale, ice stream-wide, repeat investigation would be required.

Smith *et al*. (2007) observed the creation of a drumlin over a 7 year period beneath the Rutford Ice Stream, West Antarctica using seismic investigations. Here, the seismic analysis resulted in an interpretation of erosion occurring at a rate of ~ 1 m a^{-1} for this section of the Rutford Ice Stream. This erosion then ceased and, following this, formation of a drumlin occurred from mobilised sediment at the ice-bed interface. Such mobilisation in sediments, which resulted in drumlin formation, was also attributed to hydrological changes within the glacier bed. Therefore, a combination of changes in hydrology and mobilisation of sediment was interpreted to result in drumlin formation (Smith *et al*., 2007).

Investigations of subglacial processes on active ice stream beds was also undertaken by King *et al*. (2007). Radar and seismic data from the bed of a West Antarctic ice stream identified mega-scale glacial lineations. These subglacial bedforms were interpreted to have developed in areas of dilatant, deforming till. This area of till is a section of a complex sedimentary system which is interpreted to evolve on decadal timescales. The mega-scale lineations observed beneath the modern, active ice stream were indistinguishable from ones identified on palaeo-ice stream beds. It is clear, therefore, that investigations of current ice stream processes (although rare) can allow a greater understanding of how ancient ice sheets may have behaved.

It is apparent that the lack of systematic investigations of internal composition is a result of the limited availability of sediment exposures (Stokes *et al*., 2011). The lack of well exposed drumlins limits investigations (Stokes *et al*., 2011). Furthermore, even if an individual drumlin is exposed, its sedimentological composition in relation to adjacent drumlins may not be known. Therefore, multiple sediment exposures within multiple drumlins are required for robust interpretations of subglacial processes (Figure 2.11). This problem is exemplified by Stokes *et al*. (2011), who state that ~ 40 % of papers studying drumlins report data from a small (< 5) sample from within a drumlin field.

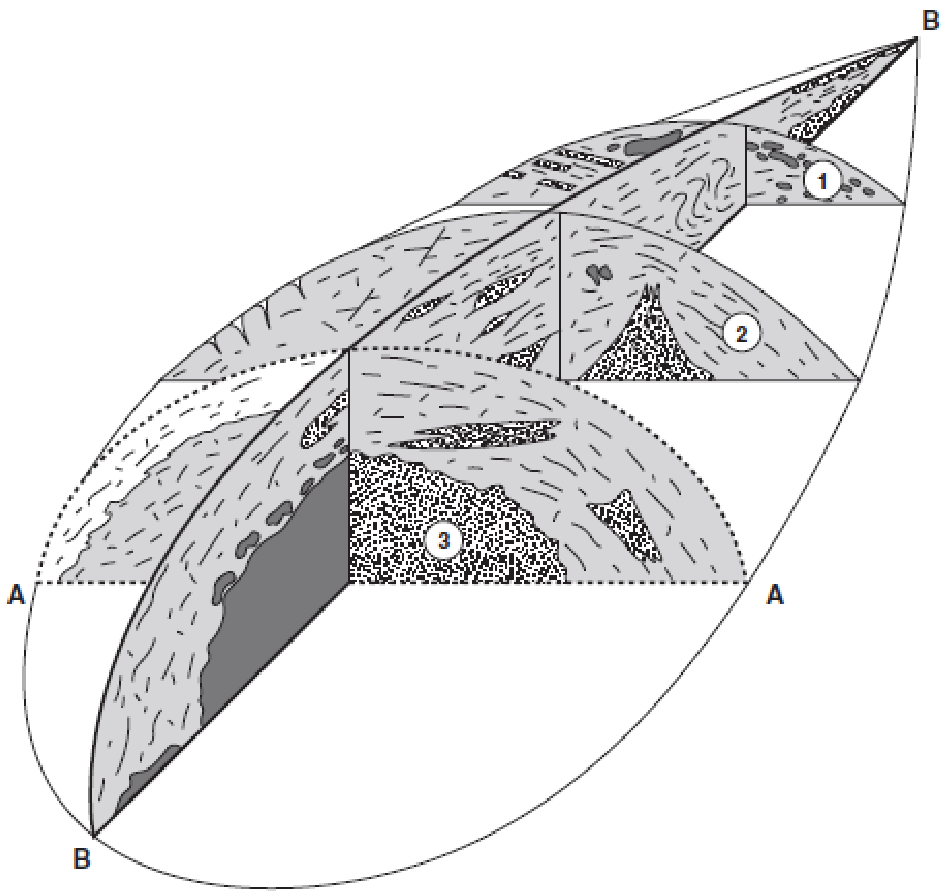


Figure 2.11. An example of the spatial complexity of the internal composition and structure within drumlins (Stokes *et al*., 2011). Investigation of sediment exposures may not allow a representative interpretation of the entire internal composition and structure of the bedform. This demonstrates the requirement for the application of geophysical analysis (Stokes *et al*., 2011).

Investigation of drumlin internal composition and structure using geophysical techniques has only recently been undertaken (Hiemstra *et al*., 2011; Spagnolo *et al*., 2014). Hiemstra *et al*. (2011) undertook seismic and electro-magnetic surveys as well as ground penetrating radar (GPR) on a drumlin in Clew Bay, Northwest Ireland. This investigation provided an entire image of the internal composition and structure (Figure 2.12), which enabled an understanding of potential processes during drumlin formation. The majority of the landform consisted of silty-clayey, deformed diamicton which was interpreted as a subglacial traction till. This unit is then surrounded by a melt-out till facies. A third, final unit, was a partly cemented, stratified sand and gravel facies which was attributed to the presence of meltwater. It was interpreted that undulating bedrock provided a nuclei for sedimentation and therefore drumlin formation. The structure of the subglacial till, occurring directly above undulations in bedrock, suggested a direct relationship between bedrock and overlying sedimentation. Further to this, clay was observed sheared into the subglacial traction till and therefore glaciotectonic processes in addition to sedimentation around pre-existing bedrock was interpreted for the creation conditions of this feature. It is evident from this investigation that the combination of sedimentary investigation and geophysical analysis provides a robust data-set for interpretations of entire bedform composition and structure.

Currently, the application of sub-bottom profiling analysis has not been considered to establish the internal composition and structure of drumlins. This is despite its widespread application to glacial sedimentology (O'Cofaigh *et al*., 2002; O'Cofaigh *et al*., 2005a; Turner *et al*., 2012; Pinson *et al*., 2013). This study, therefore, aims to demonstrate the applicability of sub-bottom profiling in alleviating problems of establishing both lateral, vertical and spatial analysis of the sub-surface composition. It is hoped that the use of sub-bottom profiling can be expanded following this study, due to the common occurrence of large areas of submerged drumlin bedforms, e.g. on lake-floors and on continental shelves occupied by palaeo-ice sheets.

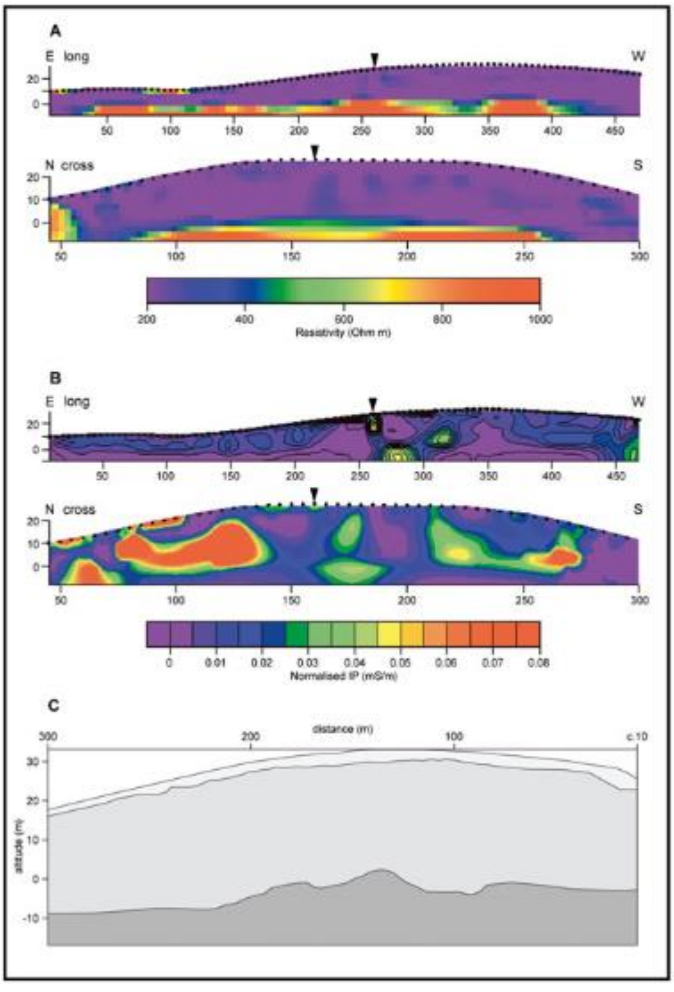


Figure 2.12. Geophysical images of a drumlin by Hiemstra *et al*. (2011) (a – c). This feature was interpreted to have formed due to sedimentation around initial till deposition, with secondary glaciotectonic processes also occurring. However, despite such studies, surveys have not been undertaken extensively to progress drumlin science, to which this study aims to contribute.

Geophysical techniques can be applied to investigate various domains within the environmental and geosciences (Reynolds, 2011). Geophysical methods can be used to extract the physical properties of the analysed materials, with such techniques allowing subsurface measurements and interpretations (Reynolds, 2011). A benefit of using geophysical techniques is that they can be applied with little environmental impact, i.e. no disturbance to the surface or subsurface (Reynolds, 2011).

Sub-bottom profiling is undertaken in this study and follows from previous applications of this technique to submerged glacial sediments (O'Cofaigh *et al*., 2002; O'Cofaigh *et al*., 2005a; Turner *et al*., 2012; Pinson *et al*., 2013). Sub-bottom profilers emit low frequency acoustic pulses which penetrate surfaces and the underlying sediment. In contrast to echo sounding investigations, acoustic energy emitted by sub-bottom profiling does not reflect off underlying surfaces. Instead it records acoustic energy reflected by layers beneath the surface. Penetration frequencies range from 1 kHz to 20 kHz with lower frequencies allowing greater depth penetration.

Sub-bottom profiling was applied by Turner *et al*. (2012) in the geophysical investigation of glacial sediments in Loch Ness, Scotland. The use of sub-bottom profiling, in association with bathymetric analysis, allowed the interpretation of a series of recessional moraines (Figure 2.13). Other glacier processes were also interpreted from the subsurface sediment composition and stratigraphic architecture of the sediments beneath Loch Ness (Figure 2.14). These include glaciotectonism occurring at a calving margin, and with iceberg and turbidite deposition occurring in the proglacial area. This example illustrates the ability to interpret individual landforms and associated sediments within the study area. These can then allow a glacial landsystem interpretation and a consideration of basin-wide glacier processes.

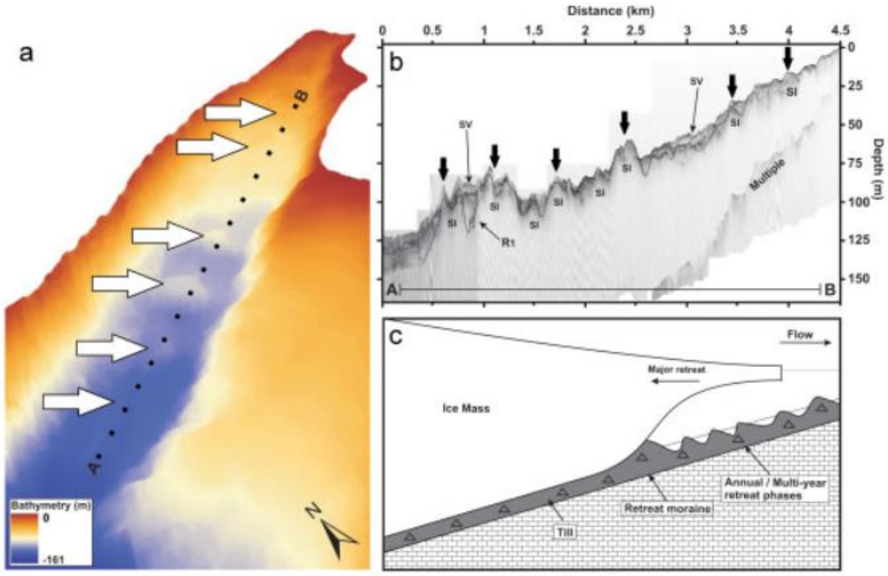


Figure 2.13. An example of bathymetric analysis (a) combined with sub-bottom profiling (b) which enabled the interpretation of former glacier dynamics (c). Annual to multi-year retreat moraines are interpreted to occur within Loch Ness and reflect recessional phases of the glacier.

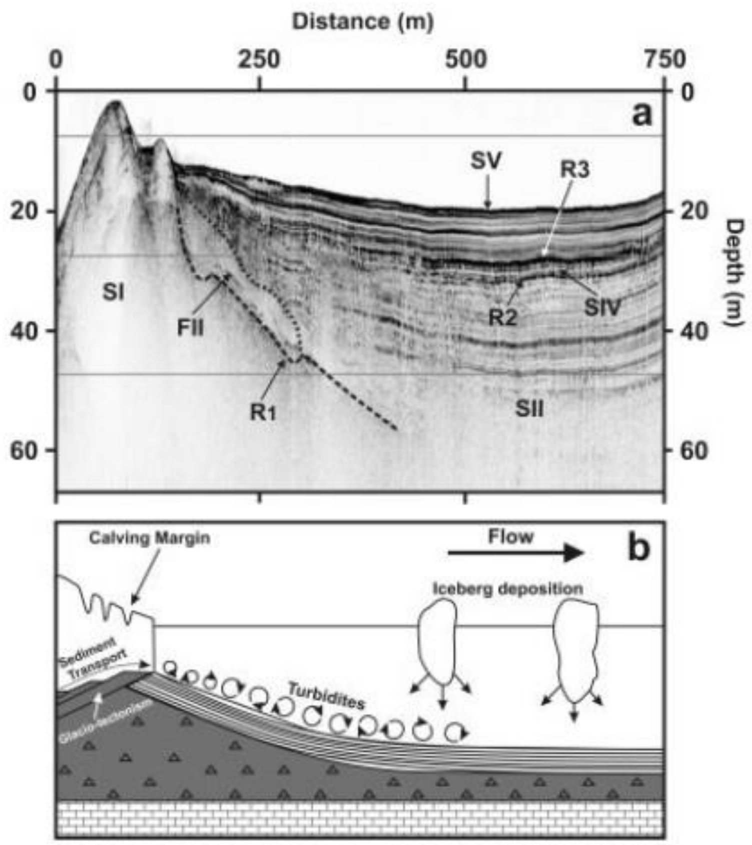


Figure 2.14. An example of former glacier processes interpreted from sub-bottom profile data (a) with a grounded ice margin calving into a proglacial lake interpreted from the sedimentary composition and structure (b) (Turner *et al*., 2012).

Sub-bottom analysis was also applied by O'Cofaigh *et al*. (2005a) to investigate the sedimentary composition of subglacial bedforms for a palaeo-ice stream that drained the Antarctic Peninsula Ice Sheet at Marguerite Bay. The sedimentary composition was then used to assess flow dynamics of a palaeo-ice stream which drained the Antarctic Peninsula. Sub-bottom profile analysis resulted in the identification of moraines, submerged by a sediment drape (Figure 2.15). Here, it was interpreted that the sediments most likely represent the grounding-zone of the palaeo-ice stream, with a subsequent sedimentation submerging the moraines. Following this, the initiation of an ice stream resulted in the deposition of a soft deformation till. Mega-scale glacial lineations were identified in a thick sequence of soft sediments on the outer shelf area and were interpreted to have formed from deforming subglacial sediments. O'Cofaigh *et al*. (2005a) therefore provided an insight into both landform specific processes and ice stream-wide processes with this illustrating the applicability of sub-bottom profiling to establishing glacial landsystems.

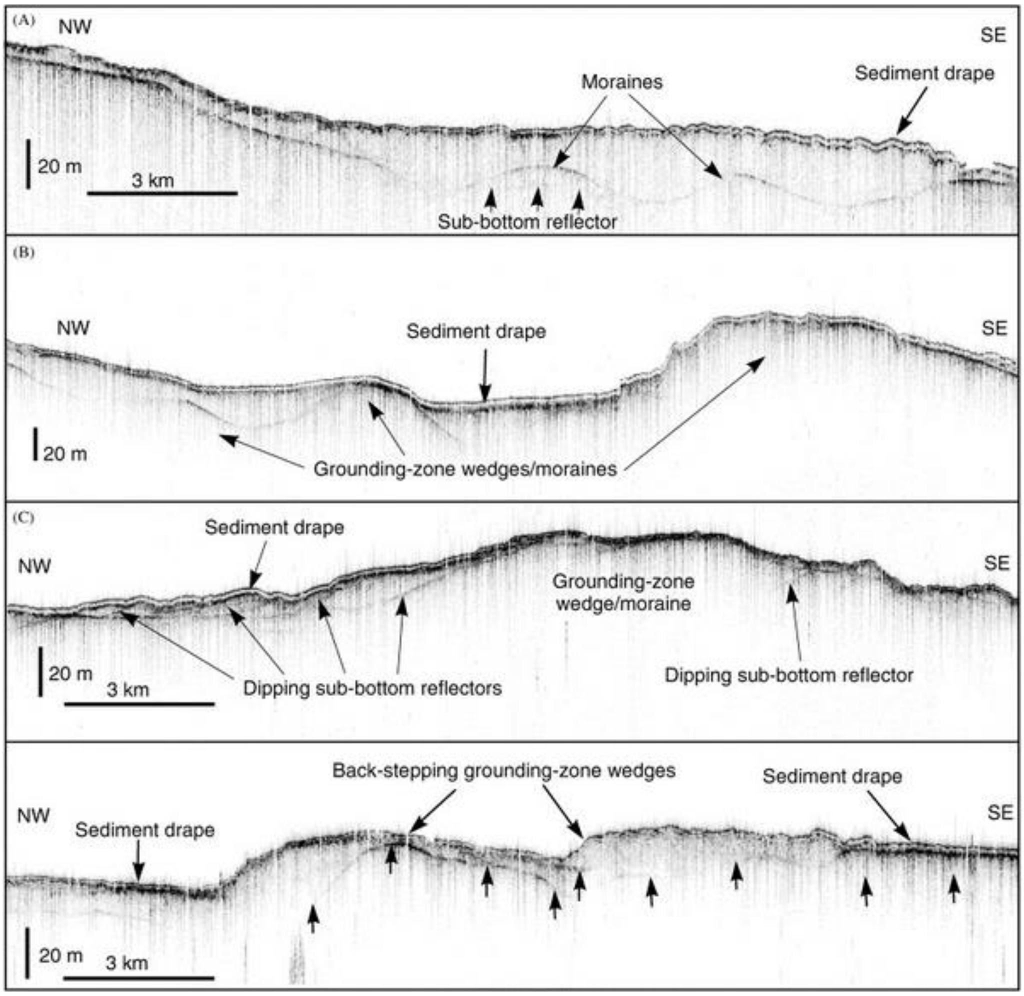


Figure 2.15. Example of sub-bottom profile data retrieved by O'Cofaigh *et al*. (2005a) for Marguerite Trough, Antarctic Peninsula, illustrating submerged bedforms by a sediment drape resulting from changes in overriding ice dynamics.

2.5. Summary

Sub-bottom sedimentary investigations have not yet been applied to establish the internal composition of drumlins that may lie in marine or lacustrine settings. This study aims to guide future large-scale investigations of internal drumlin composition and structure through testing the use of sub-bottom profiling in a lacustrine setting where drumlins are thought to be present. Due to the abundance of drumlins within marine and lacustrine settings, this technique has the potential to reveal vast swathes of sub-surface composition and structure. Through such analysis a greater understanding of ancient subglacial processes is hoped to be established.

3. STUDY SITE

3.1. Lake District glaciations and Bassenthwaite Lake

The Last Glacial Maximum (LGM) of the British-Irish Ice Sheet occurred at ~ 26 – 22 ka BP (Boulton and Hagdorn, 2006; Hubbard *et al*., 2009; Clark *et al*., 2012). Recent reconstructions suggest ice streams were active between 24 – 19 ka BP with a large-scale re-organisation of ice sheet flow occurring during this period. Recession is interpreted to have occurred over a relatively short period of ~ 4 ka, between 19 – 15 ka BP. This resulted in large-scale deglaciation and a transition from large-scale ice sheet and ice stream processes, to more localised, regional valley and glacier deglaciation (Figure 3.1).

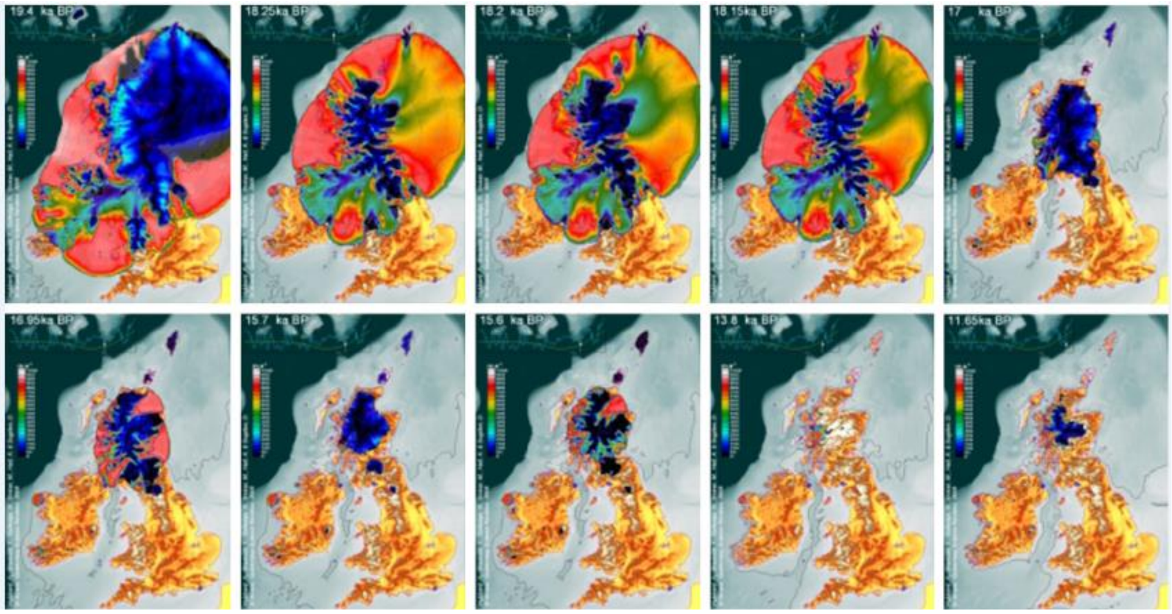


Figure 3.1. Illustration of deglaciation of the British-Irish Ice Sheet as interpreted through forward transient numerical modelling by Hubbard *et al*. (2009). The extent of the British-Irish Ice Sheet from 19.4 ka (top left) to 11.6 ka (bottom right) is illustrated with the addition of blue shaded areas showing frozen basal conditions and areas of red showing mobile basal conditions. Within the Lake District, overall ice mass processes would have changed from large-scale ice stream and ice sheet subglacial conditions to more localised valley glacier recession.

Ice stream initiation and dynamics were investigated by Evans *et al*. (2009) for the central sector of the British and Irish Ice Sheet, with this area of the British-Irish Ice Sheet the focus of recent investigations (Livingstone *et al*., 2008; Evans *et al*., 2009). Here, ice stream flow phases are categorised for the northern Lake District, Solway Firth, Tyne and Stainmore Gap. The first flow phase is considered to have originated from Scottish dispersal centres and forced Lake District ice eastwards over the Pennines. The second phase results in a continued easterly flow of Lake District ice. Finally, a third phase results in a switch of ice streaming to the west with ice drawn down through the Solway Firth and into the Irish Sea Basin (Figure 3.2; Figure 3.3).

Ice stream occurrence is recorded in the north of the Lake District. Here, ice flowed into the north Irish Sea Basin, as observed through geomorphological mapping of subglacial features (Figure 3.3; Livingstone *et al*., 2008). Differentiation between ice stream flow phases is also provided with cross-cutting evidence for dynamic ice flow. This is interpreted to have occurred as a result of alterations in ice dispersal centres (Evans *et al*., 2009).

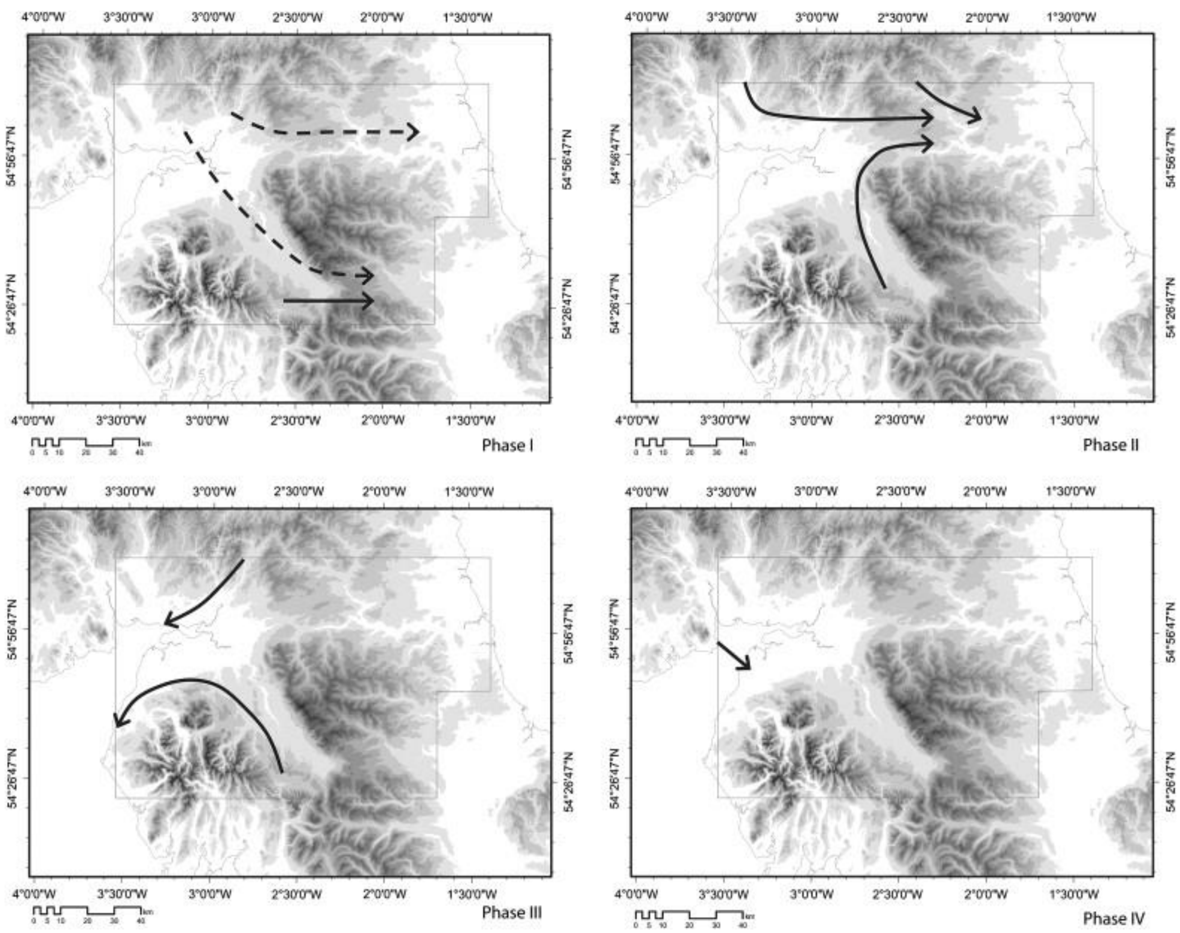


Figure 3.2. Summary of the principal ice flow events for the central sector of the British and Irish Ice Sheet illustrating a generalise switch from east, to west ice stream drainage (Evans *et al*., 2009).

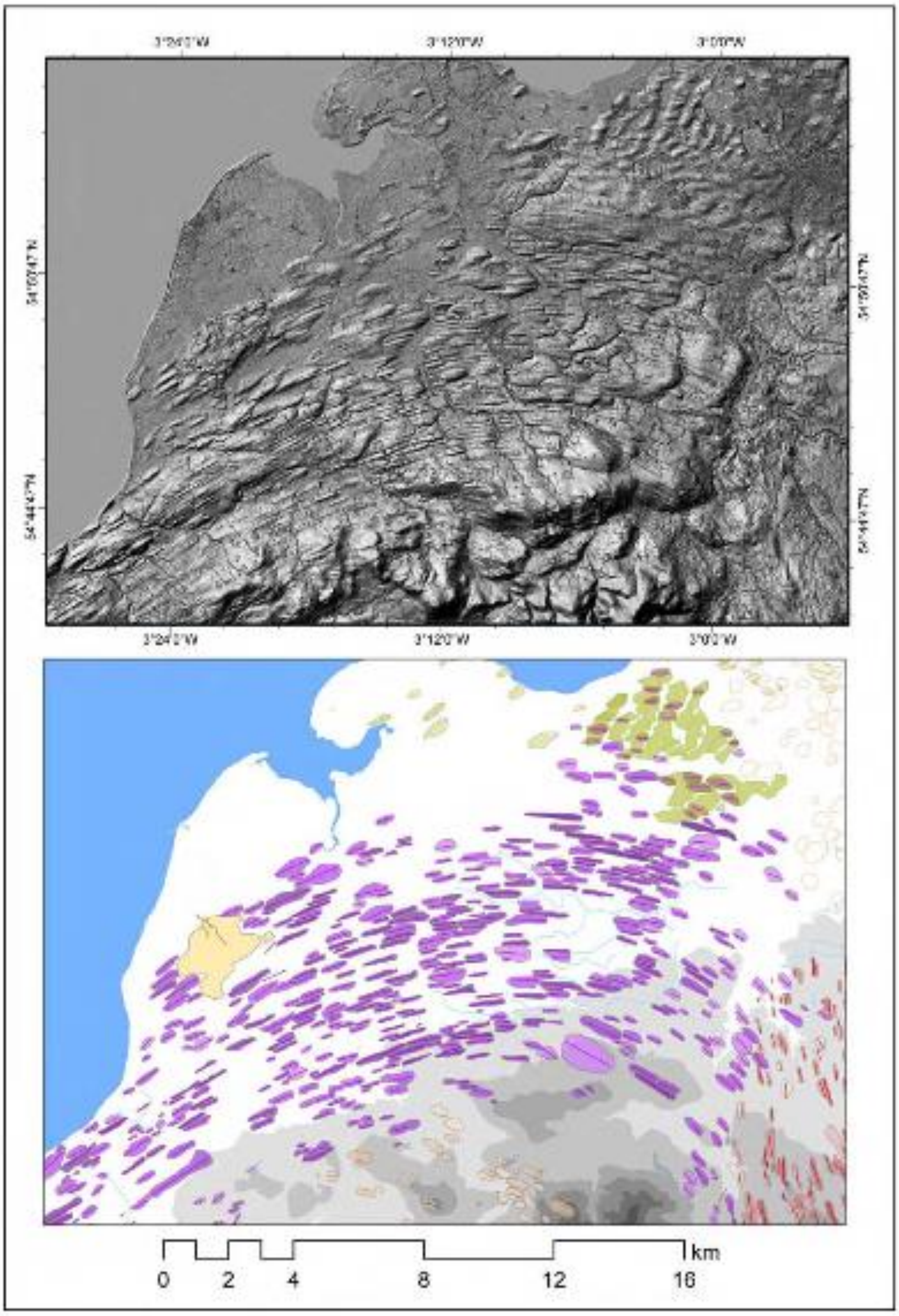


Figure 3.3. Subglacial geomorphology recorded by Livingstone *et al*. (2008), illustrating ice flow around the north of the Lake District and into the north Irish Sea Basin. This is illustrated by Evans *et al*. (2009) to represent a transition in ice stream flow, which occurred initially from west to east.

A dearth of knowledge exists regarding ice sheet retreat dynamics in the Lake District. Deglacial, sedimentary investigations are rare across the Lake District and most work has focussed on the Loch Lomond Stadial (Younger Dryas). However a few late glacial sedimentological investigations have been undertaken (Ballantyne *et al*., 2009; Pinson *et al*., 2013). Deglaciation dynamics have been considered for Lake Windermere (Pinson *et al*., 2013). In Lake Windermere, a combination of geophysical analysis and geomorphological investigation allowed the observation of recessional moraines, over-consolidated till, De Geer moraines, ice-frontal fans, supra-glacial debris and outwash material (Figure 3.4; Pinson *et al*., 2013). The composition and distribution of these features were interpreted to reflect a dynamic, valley glacier with downwasting and ice-stagnation resulting in a series of complex supra-glacial and glacio-lacustrine deposits. However, Pinson *et al*. (2013) also interpret an active ice margin occurring within Lake Windermere, prior to ice stagnation and downwasting. Such deglacial dynamics observed within Lake Windermere provide an insight into former glacier processes for other glaciated areas within the Lake District.

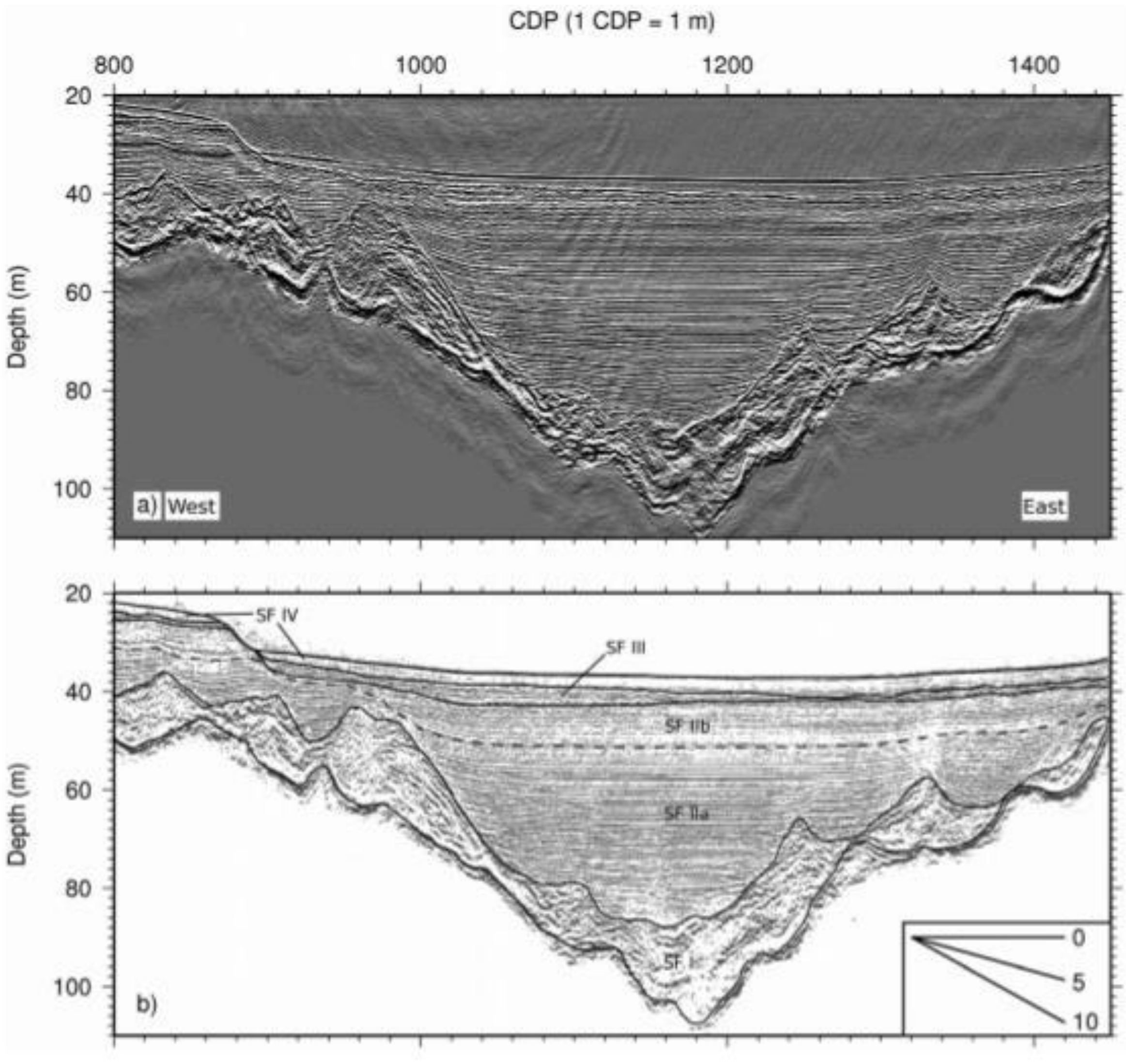


Figure 3.4. Geophysical profile running west to east at a southern basin of Lake Windermere. Here, Pinson *et al*. (2013) interpret a glaciolacustrine infill resulting from distal deposition and a receding glacier.

Detailed, basin-specific geomorphological and sedimentological investigations of Bassenthwaite Lake and adjacent areas are absent. General geomorphology adjacent to Bassenthwaite Lake, however, has been mapped by Livingstone *et al*. (2008) and Hughes *et al*. (2010). These investigations were undertaken to establish large-scale ice stream and ice sheet bedforms as opposed to localised, valley glacier retreat. Hughes *et al*. (2010) recorded numerous drumlins, both adjacent to Bassenthwaite Lake, and in the valley basin leading towards Derwent Water (Figure 3.5; Hughes *et al*., 2010). However, no sediment exposures from these drumlins have been reported.

Following ice sheet, ice stream and valley glacier deglaciation, a return to colder conditions during the Younger Dryas resulted in the creation and advance of numerous ice masses within the Lake District (Brown *et al*., 2013; Figure 3.6). Although ice did not advance into Bassenthwaite Lake directly, the environmental change associated with this likely resulted in changes in sediment input into Bassenthwaite Lake.

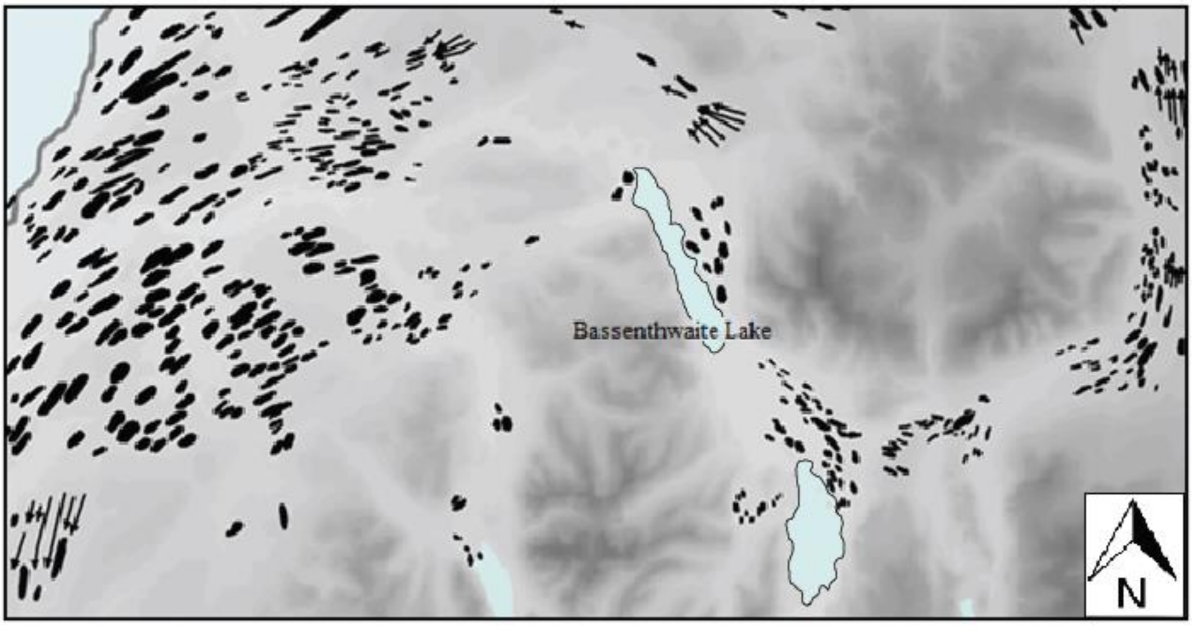


Figure 3.5. Glacial geomorphology of the north Lake District with Bassenthwaite Lake labelled. Black features are subglacially rounded bedforms (drumlins) with orientations provided from some bedforms (Hughes *et al*., 2010). Terrestrial drumlin features adjacent to Bassenthwaite Lake provide evidence for potential submerged drumlin features.

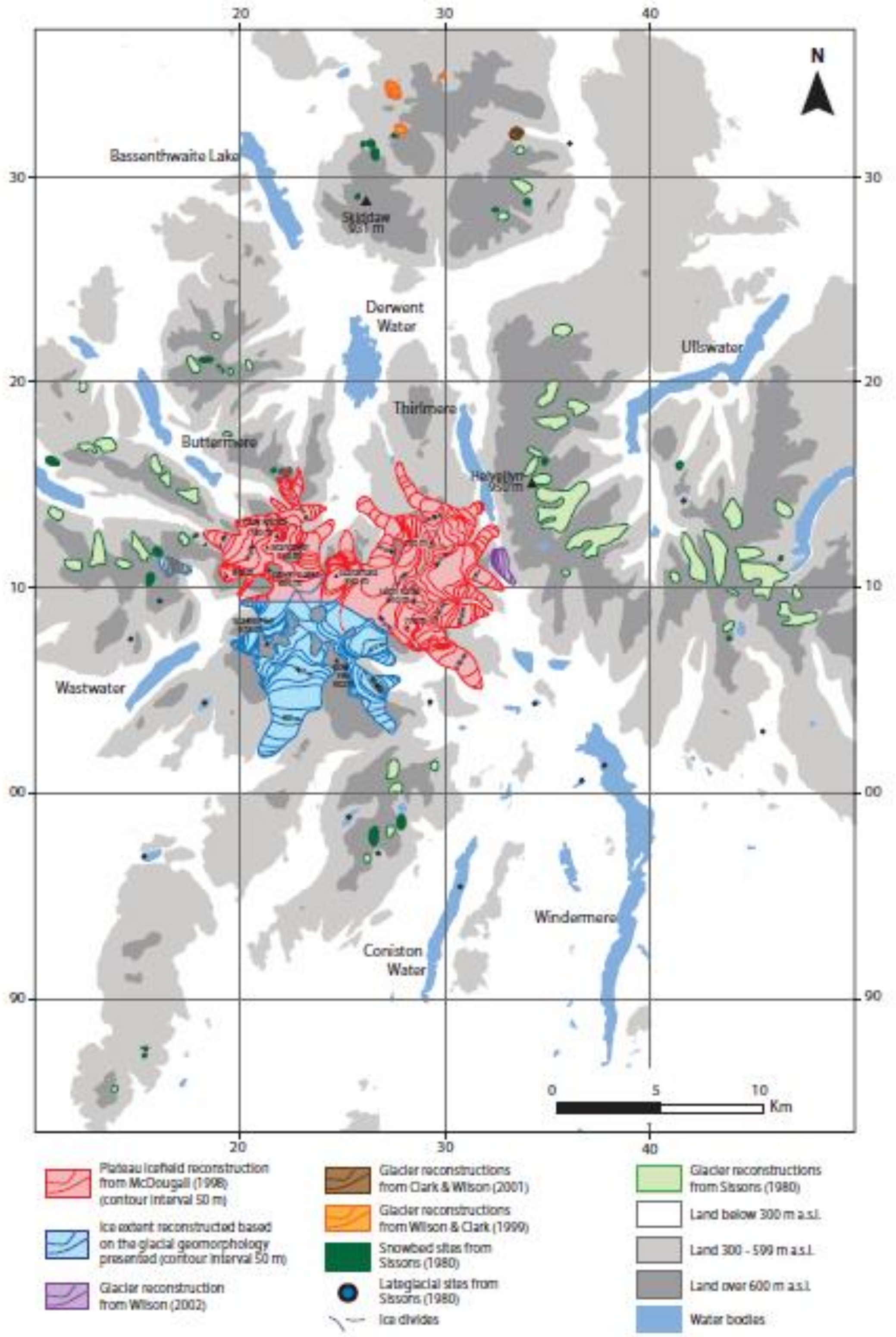


Figure 3.6. Younger Dryas ice extent compiled by Brown *et al*. (2013) for the Lake District. Such Younger Dryas glacier locations provide evidence for deglacial dynamics and potential ice cap accumulation processes for prior glaciations.

3.2. Bassenthwaite Lake

3.2.1. Justification for selection of Bassenthwaite Lake

Bassenthwaite Lake was selected for both geophysical and geomorphological investigation due to the potential for both submerged subglacial bedforms and for lake-adjacent features. A very simple bathymetrical map provided by Ramsbottom (1976) suggests subglacial features may be present on the lake floor. Furthermore, drumlins were mapped adjacent to the lake by Hughes *et al*. (2010). Therefore, the key justification for the selection of Bassenthwaite Lake is that this area, in addition to Derwent Water, provide some of the only locations across the entire Lake District where drumlins have been mapped adjacent to lakes and probably lie on the lake floor, making them suitable targets for geophysics. The location of Bassenthwaite Lake, recording deglacial dynamics of the area, was also an influential factor when selecting study sites. Palaeo-ice flow from the Bassenthwaite basin into the Solway Lowlands has resulted in both large-scale ice stream and local valley glacier processes recorded in the geomorphic-sedimentary record. Access to lakes across the Lake District was also of consideration, with a scientific permit required for access to Bassenthwaite Lake. Derwent Water (the nearest water body to Bassenthwaite Lake) does not easily permit boat access although numerous drumlin features were also identified both within and adjacent to that more southerly lake. The absence of any anthropogenic activities within Lake Bassenthwaite also provides justification for undertaking sub-bottom analysis at this location.

3.2.2. Background information, topography and geology

Data collection was undertaken at Bassenthwaite Lake, in the north of the English Lake District (Figure 3.7). Bassenthwaite Lake is the most northern and shallowest lake in the English Lake District. The Lake has a mean depth of 5.3 m, a deepest point of ~ 20 m and an average altitude at 335 m (Thackery *et al*., 2010). The lake has three main inflows which include the River Derwent, Newlands Beck and Chapel Beck. The River Derwent provides 80 % of the lake's hydraulic input and drains the eastern area of the Lake's catchment. The catchment of Bassenthwaite Lake includes Derwentwater and Thirlmere with these two water bodies effectively acting as upstream sediment traps. Topography adjacent to Bassenthwaite Lake varies greatly from > 900 m at Skiddaw, to the outflow of Bassenthwaite at 71 m. The catchment of Bassenthwaite Lake is dominated by the Skiddaw Slate Group (metamorphosed Ordovician, marine turbiditic and hemi-pelagic greywacke, arenite, siltstone and mudstones) (British Geological Survey, 1999; Figure 3.8). This Skiddaw Slate Group also underlies Bassenthwaite Lake. However, small areas of Borrowdale Volcanic Group (extrusive volcanic association) are apparent in the southwest of the catchment. Palaeo-ice flow would have occurred from the south of the lake basin towards the north of the lake, opening into the Solway Lowlands (Evans *et al*., 2009).

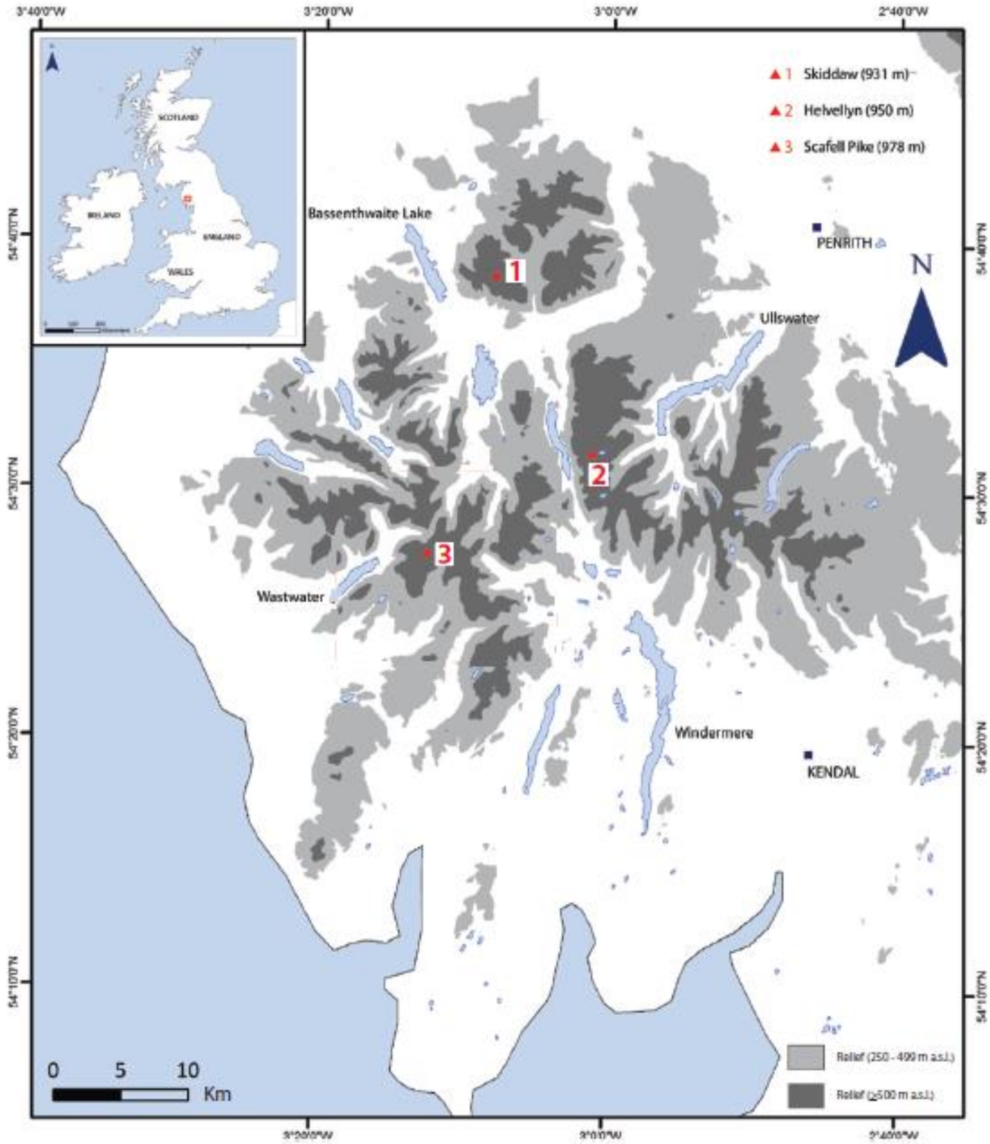


Figure 3.7. Location of Bassenthwaite Lake in relation to both the UK (inset) and the entire English Lake District (modified from Brown *et al*., 2013). Here, the three highest peaks are also mapped with Bassenthwaite Lake adjacent to Skiddaw (931 m).

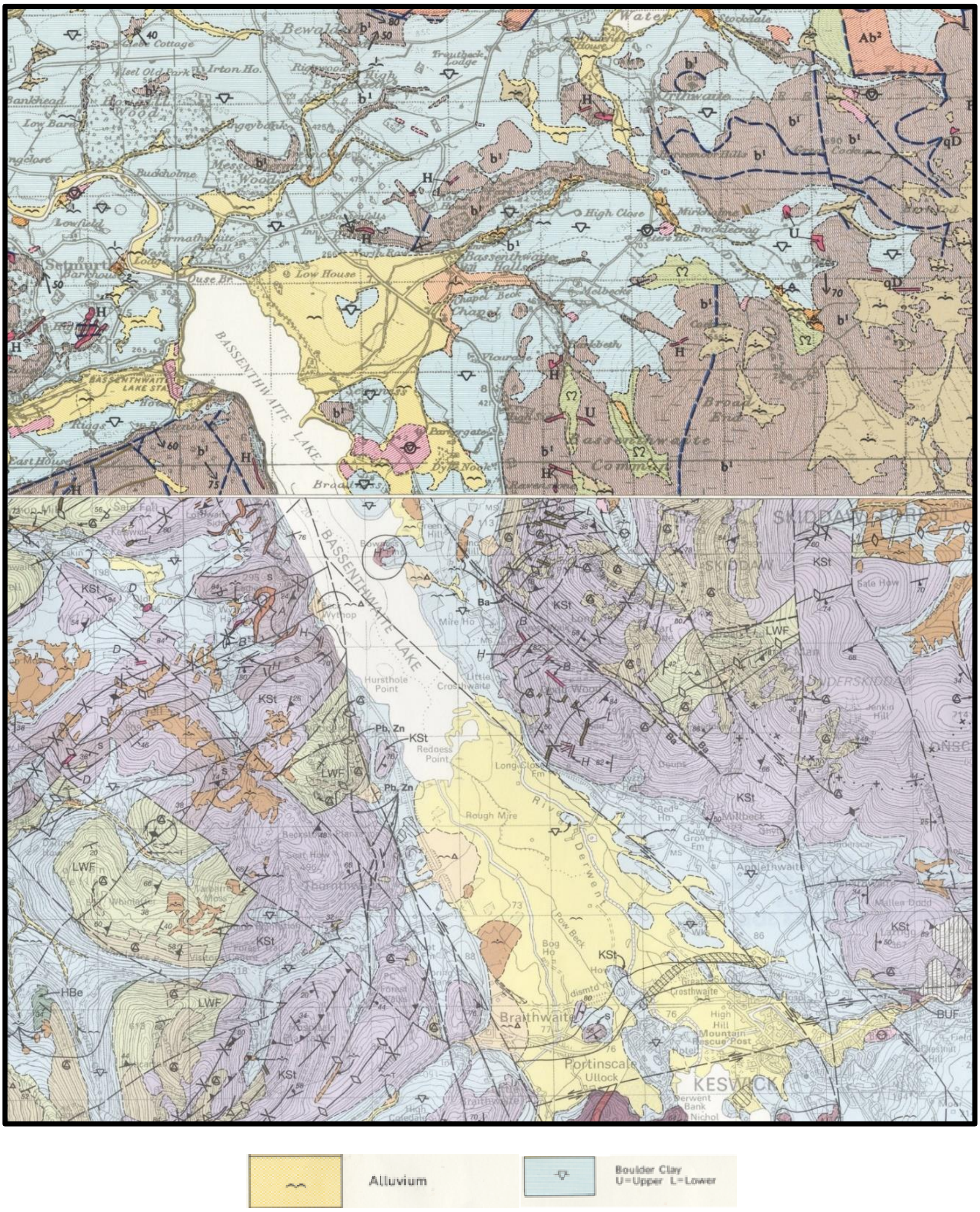


Figure 3.8. Bedrock geology and drift for Bassenthwaite Lake and adjacent areas, dominated by: Skiddaw Slate Group (purple areas), alluvial drift and boulder clay (British Geological Survey, 1999).

3.2.3. Lake Bathymetry

A small number of bathymetry profiles have been provided from Bassenthwaite Lake from the geophysical analysis of Ramsbottom (1976) (Figure 3.9). However, Bassenthwaite Lake currently lacks a high resolution bathymetric dataset. Such investigations have been undertaken for other Lake District lakes, including Lake Windermere (Figure 3.10). These have provided insights into ice dynamics, deglaciation processes and postglacial sedimentation (British Geological Survey, 2012).

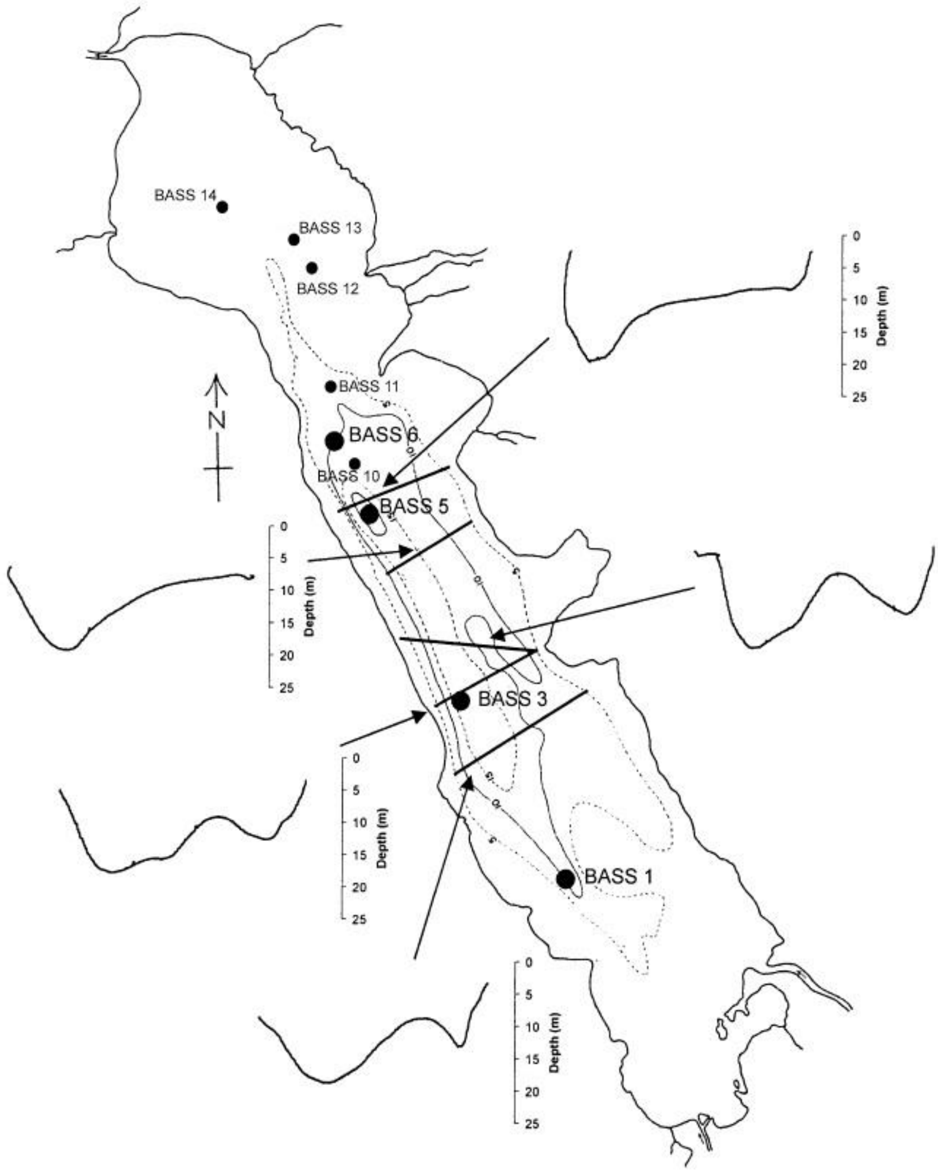


Figure 3.9. Bathymetry of Bassenthwaite Lake as produced by Ramsbottom (1976). Lateral bathymetry profiles are also provided and were undertaken through geophysical analysis. These illustrate potential locations to investigate submerged drumlins.

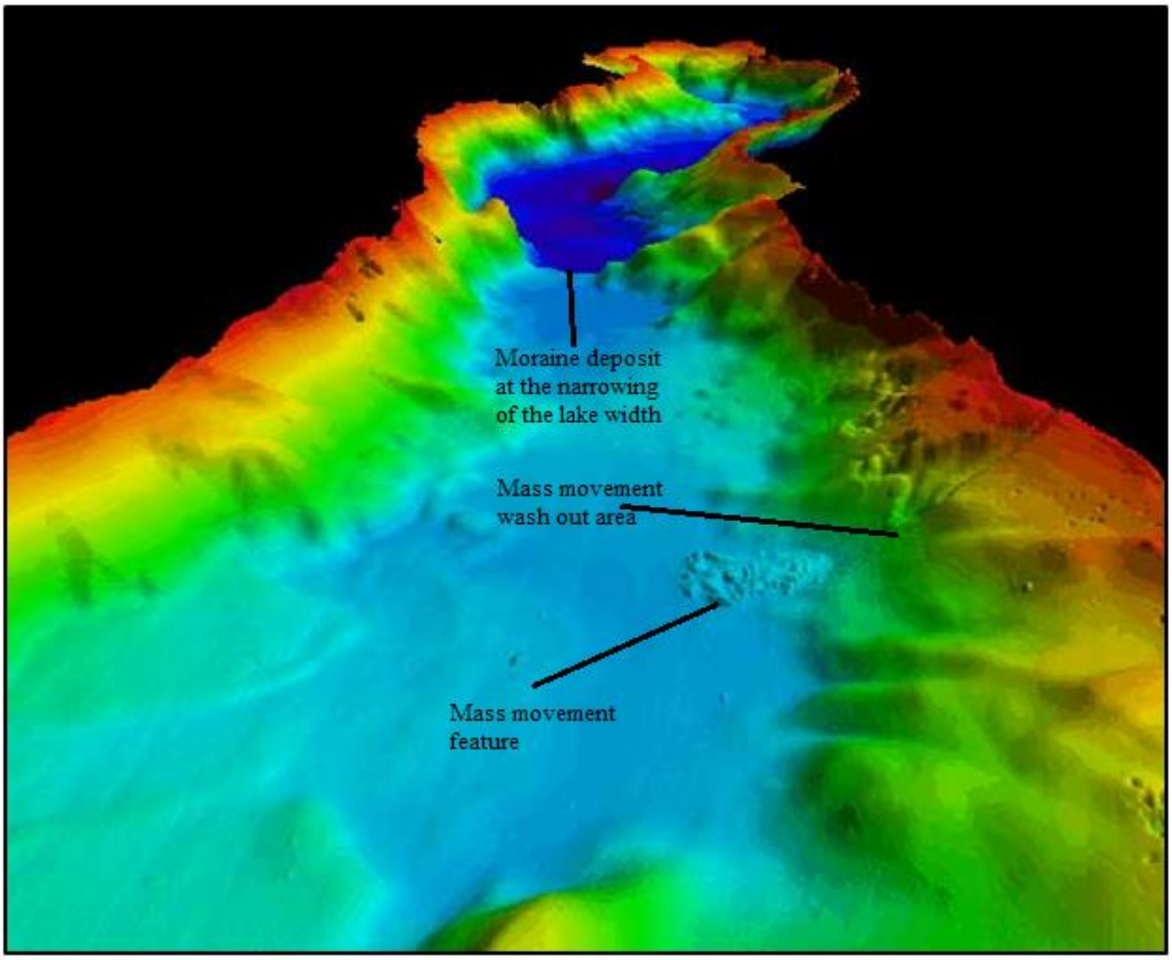


Figure 3.10. Bathymetry of Lake Windermere as investigated through multibeam geophysical analysis by the British Geological Survey. Multibeam analysis contributed towards interpretations of the glacial landsystems of Lake Windermere. However, mass movement events associated with postglacial environmental change were also discovered. This allowed a consideration of the environmental history following deglaciation (British Geological Survey, 2012).

4. METHODOLOGY

4.1. Introduction

The primary focus of this study is the analysis of lake-floor geomorphology and its sub-surface composition and structure using geophysical, sub-bottom profiling. In addition to this, geomorphological mapping was undertaken to identify features adjacent to Bassenthwaite Lake. Sub-bottom profiling was undertaken on Bassenthwaite Lake from the 19th to the 21st of May, 2014. Geophysical equipment (EdgeTech 3100P Sub-bottom Profiling System with a SB-424 towfish (Figure 4.1)) was provided by Northumbria University. The application of this type of geophysical analysis circumvents intrusive data collection and is commonly used in formerly glaciated areas to investigate ice sheet-sedimentary processes (O'Cofaigh *et al*., 2002; Hogan *et al*., 2011; Turner *et al*., 2012) (see section 2.4). An investigation of geomorphology through mapping and geophysical analysis can allow the identification of sediment-landform associations and enable glacial landsystem reconstructions. Possible submerged subglacial landforms have been previously identified from low resolution lake bathymetry from Bassenthwaite Lake (Ramsbottom, 1976; Figure 3.9). However, large swathes of the lake floor were re-surveyed to provide a record of the lake floor bathymetry and sub-bottom stratigraphy.

Figure 4.1. The SB-424 Sub-bottom profiling 'towfish' that is pulled through the water column behind the towing vessel.

4.2. Geomorphological mapping

This study provides the first detailed glacial geomorphology maps for Bassenthwaite Lake and the adjacent area. The geomorphology at Bassenthwaite Lake and adjacent areas was interpreted from a Digital Elevation Model (DEM) sourced from NEXTMap Britain and all features were digitised electronically and stored as shapefiles. Google Earth^tm imagery was also used to identify landforms. The initial mapping from DEMs and satellite images was ground truthed during geophysical data collection from the 19th to the 21st of May, 2014. The UK-based glacial geomorphological mapping from Hughes *et al*. (2010) was also incorporated within the geomorphological map (Figure 4.2). The geomorphological map produced here covers an area of ~ 7 x 9 km (Figure 5.1). Rivers flowing into Bassenthwaite Lake at both the southern and northern ends of the lake basin were mapped as they have the potential for sediment input which could be recorded within sub-bottom transects. The valley basin area between Bassenthwaite Lake and Derwent Water was also mapped due to its importance in reflecting deglacial geomorphology similar to that submerged within the lake. Specific criteria were applied to identify and digitise landforms, as outlined below:

4.2.1. Drumlins

Drumlins (streamlined, oval features with the long axis parallel to the orientation of ice flow) were identified as changes in topography and the presence of oval shaded features from aerial images and were mapped following criteria outlined by Stokes and Clark (2001), Stokes and Clark (2002) and Spagnolo *et al*. (2010). The drumlins were mapped as polygons around the break of slope. Drumlins occur frequently within the mapped area as clustered oval-shaped features with average lengths of ~ 400 m and some longer bedforms occurring to ~ 700 m. These were also identified occurring parallel to potential inferred ice flow and located adjacent to other drumlin features (owing to their likelihood to occur in swarms). The morphology, distribution and size of drumlins are of importance as they may illustrate changes in subglacial processes spatially across the glacier bed. Features of this morphology were also mapped by (Mitchell and Riley, 2006) in the Western Pennines and were also interpreted as drumlins forming part of a larger drumlin field. Outside of the UK features of this morphology were also interpreted as drumlins, using the same criteria as those used here, by Darvill *et al*. (2014) in southern Patagonia, where they occur in association with other glacial lineations.

4.2.2. Moraines

These features are linear to curvilinear ridges, occurring perpendicular to potential palaeo-ice flow. Moraine features appeared through direct field investigation and through remotely sensed images as linear ridges of alterations in relief and were mapped using polygons around the break in slope.

Moraines where identified on DEMs as changes in contour lines and on Google Earth[tm] as shaded linear ridges. Typically the mapped moraines are ~ 300 m in length and ~ 5 – 8 m in height and clearly run in to Bassenthwaite Lake. The moraine ridges occur both continuously (for up to ~300 – 400 m) and discontinuously (~ 30 m) throughout the Bassenthwaite Lake basin. These features are primarily identified through their linearity, perpendicular across the basin of the valley. Such continuous and discontinuous ridges resulting in a positive relief and occurring perpendicular to ice flow have been interpreted in numerous investigations as moraine ridges (Clark *et al*., 2004; Darvill *et al*., 2014).

4.2.3. Gullies

Discrete and sinuous areas of a decrease in topography and an anabranching, dendritic pattern are interpreted as fluvial incised gullies and ephemeral streams. These features are located within topographic lows and merge into larger channels. Gullies are illustrated within DEMs and satellite imagery as narrow (~ 2 m) decreases in elevation, resulting from fluvial incision. These gullies then occur as sinuous channels, increasing in width down-slope with the addition of feeding gullies. Features of this morphology were also mapped by Sahlin and Glasser (2008) for the Cadair Idris area, Wales, and were also interpreted as fluvial incision features which occurred following glacier retreat.

4.2.4. Mass-movement features

Areas of unconsolidated accumulations of material were identified from alterations in shading within satellite imagery. The accumulation of such unconsolidated material also occurs beneath high altitude peaks are interpreted as scree-slope material. These features primarily occur to the east of Bassenthwaite Lake along the high altitude (~ 900 m) areas of Skiddaw. Features of this morphology were also mapped by Sahlin and Glasser (2008) and interpreted as scree slope material resulting from post-glacial, sub-aerial weathering for the Cadair Idris area, Wales, UK.

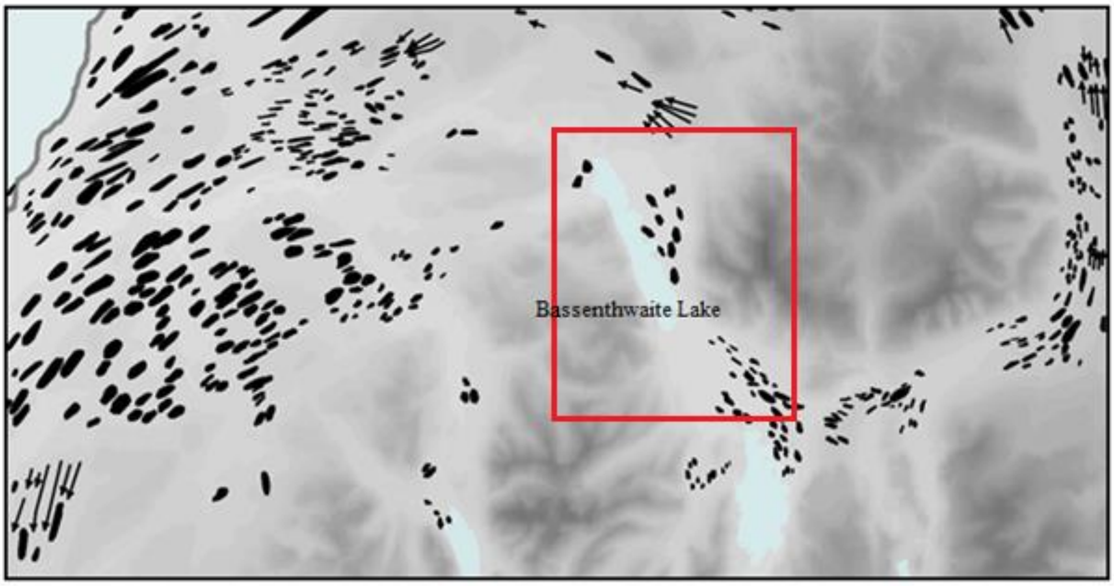


Figure 4.2. Location of study area for geomorphological mapping superimposed upon mapping of subglacial features undertaken by Hughes *et al*. (2010).

4.3. Sub-bottom profiling

4.3.1. Principles

Sub-bottom profilers emit low frequency acoustic pulses which penetrate bounding surfaces and underlying sediment (Figure 4.3). In contrast to echo sounding systems, acoustic energy emitted by sub-bottom profiling does not reflect off underlying surfaces. Instead, it records acoustic energy reflected by layers beneath the surface. Penetration frequencies range from 1 kHz to 20 kHz with lower frequencies allowing greater penetration depth (up to 150 m). The sub-bottom profiler is towed behind a boat and is connected by a steel cable and an electric cable (responsible for transmitting the collected data to the processing unit (Figure 4.4)). The depth of the profiler within the water column can be altered depending on the necessity for specific data collection (for example, lowering the profiler within the water column can provide greater penetration depths). The electric cable runs from the submerged sub-bottom profiling 'tow fish' to a processing unit housed within the boat. The processing unit then transfers the collected data to a laptop computer. This displays the reflective properties of the materials analysed through Discover SB 3200-XS software. GPS then provides coordinates for the collected data to allow the spatial referencing of the data (Figure 4.3).

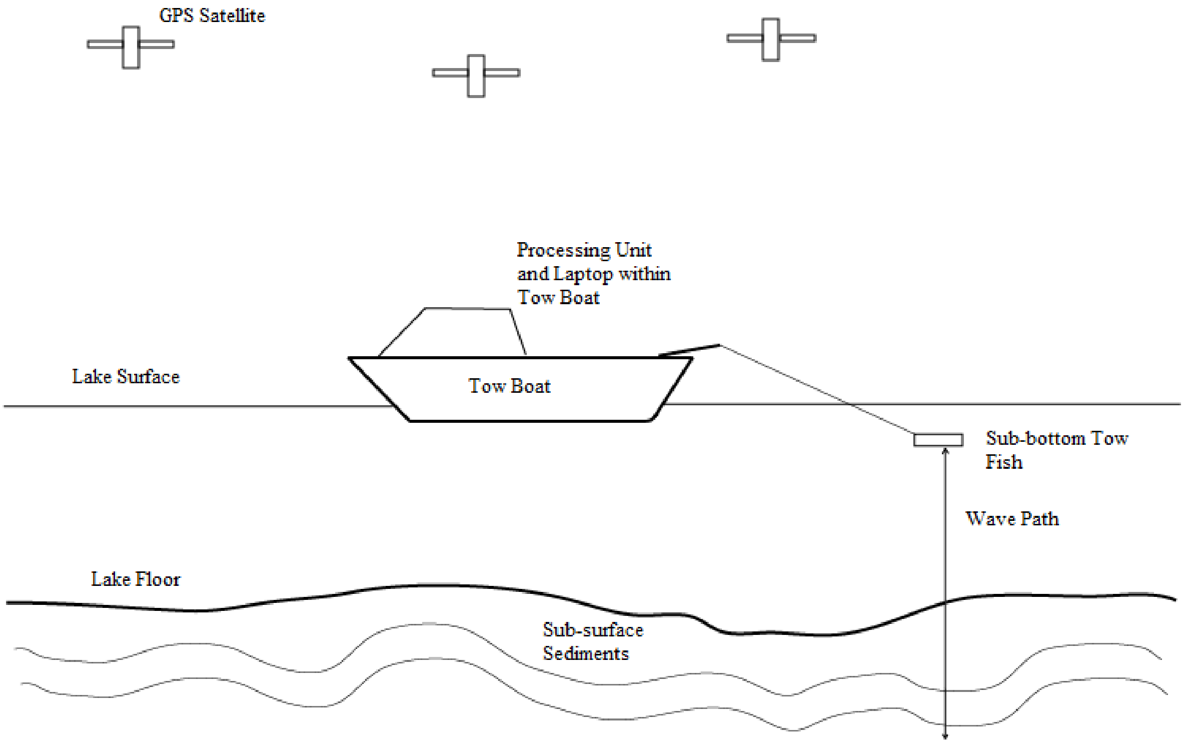


Figure 4.3. Schematic of the components considered during sub-bottom profiling.

Figure 4.4. 565P Topside processing unit which records all data collected from the attached towfish and transfers data to the laptop unit for visual analysis.

4.3.2. Data acquisition

Sub-bottom profiling was undertaken during traverses of the SB-424 towfish (Figure 4.1) which was attached to a small boat. The towfish profiler was set to varying depths (2 and 4 metres) dependent upon proximity to the lake floor and in order to optimise detailed data collection. For example, lowering of the tow fish to 4 metres resulted in increased detailed data acquisition, which was hypothesised to result from the increased proximity of the tow fish to the lake bed. The towfish operated at a pulse frequency of 20 kHz which was established through the experimentation of varying pulse lengths upon the sedimentary packages under investigation. An initial application of 15 kHz resulted in a 'bleached' and 'faded' image displayed through the Discover SB 3200-XS software. The change to 20 kHz resulted in the observation of high resolution, stacked sedimentary package. The Discover SB 3200-XS software allows the recording of sub-bottom data with transects produced as .jpg files which were later visually interpreted. Sub-bottom data from within Bassenthwaite Lake varied spatially with some areas revealing detailed sedimentological structures and other locations producing no apparent detail (Figure 4.5). Alterations in surface sediment chemical composition and changes in lake-floor vegetation may be responsible for spatial variations in the signal.

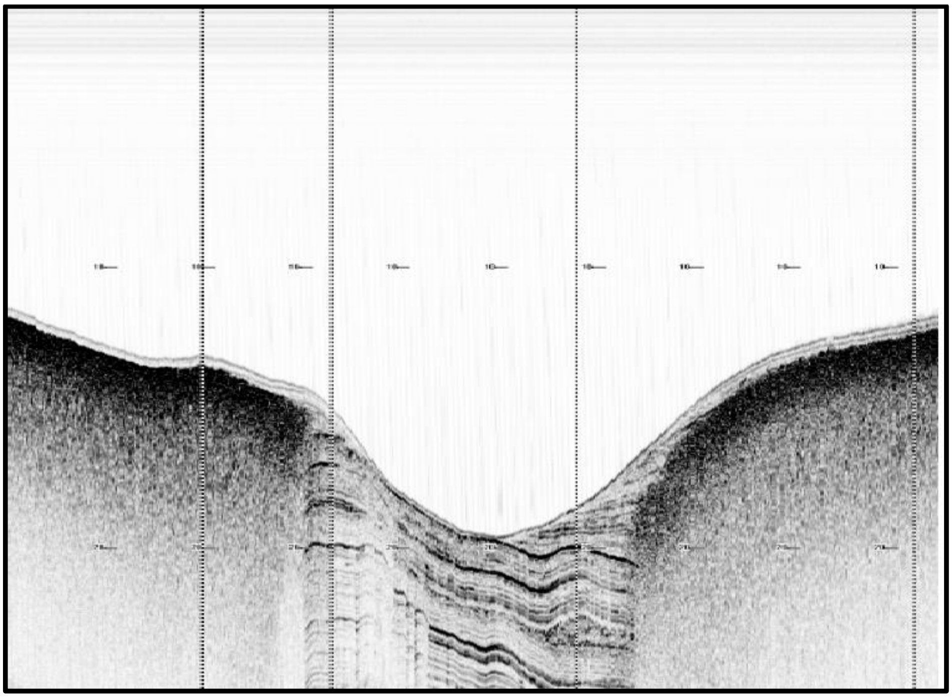


Figure 4.5. Example of the spatial variation in sub-bottom composition detail. The areas with a thin sediment drape, followed by a reflected black layer that fades gradually into grey represents the areas where acoustic penetration did not occur beyond the thin (~ 50 cm) drape. Detailed acoustic reflections occur within the middle of the image as apparent by numerous reflected sedimentary layers. The spatial variations in detailed imagery have been interpreted as a result of changes in sub-bottom properties and not large-scale variations in sedimentary composition and structure.

4.3.3. Transect locations

Collection of sub-bottom data was undertaken through traversing Bassenthwaite Lake with the sub-bottom profiler. Initially, discrete areas were outlined for geophysical investigation which were believed to contain submerged drumlins, based on previous bathymetric profiles. Transects were undertaken from the north to the south of the lake (along the length of the lake), with ~ 10 m changes in distance to ensure entire lake coverage was undertaken. When detailed areas were obtained and observed in 'real time' using geophysical software, the longitudinal transects were then supplemented with traverses across the width of the lake to provide lateral context. From such traverses, transects containing detailed sedimentology were established (Figure 4.6). A possible submerged drumlin identified on bathymetry maps prior to fieldwork provided no detailed sub-surface sedimentology. The transects provided within Figure 4.6 do not represent all locations investigated, but instead illustrate transects which contained detailed sedimentological information.

4.3.4. Post-fieldwork data processing, interpretation and presentation

Geophysical, sub-bottom images were recorded using the Discover SB 3200-XS software onto a field-laptop. These are recorded as .jpg images and were transferred onto an external hard-drive. The .jpg images were then manipulated through Adobe Illustrator. This includes the cropping of transects where areas of no detailed sediments occur. Following this, an overlay of interpretations of the individual lithofacies and stratigraphic architecture was undertaken. Lithofacies Associations were then identified and grouped. This allowed a spatial comparison of sedimentological properties between the detailed sub-bottom images. Individual lithofacies were grouped by assessing numerous internal characteristics of the .jpg images. These included the strength of reflectance of the stratigraphic units, in addition to the thickness of bedding, presence of stratification, presence of deformational features and thickness of each individual unit. This discrimination between sedimentary units and the allocation of lithofacies then allowed interpretations of depositional environments and an interpretation of past ice processes from the structure of such lithofacies.

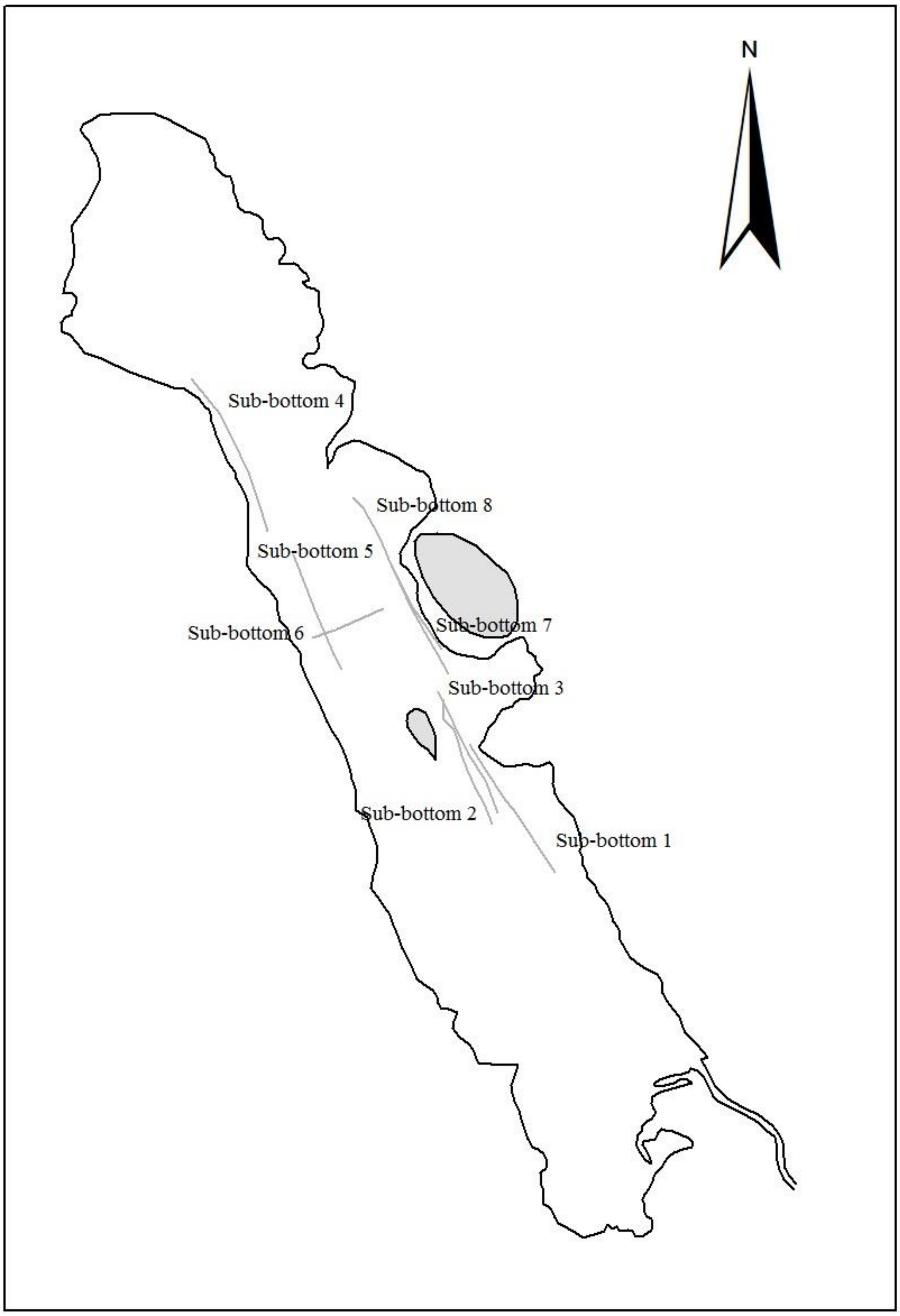


Figure 4.6. Bassenthwaite Lake, with submerged and lake-adjacent drumlin landforms (grey, filled features). Here the grey lines represent transects containing detailed sub-bottom sedimentological information.

5. RESULTS

5.1. The glacial geomorphology of Bassenthwaite Lake and adjacent areas

Bassenthwaite Lake is situated in the northern Lake District. It is ~ 6 km long, ~ 1 km wide and ~ 20 m deep. It occupies the floor of a lowland trough (~ 70 m OD) running SE to NW between the mountainous areas of Skiddaw (931 m OD) and Broom Fell (511 m OD). The geology of this sector of the Lake District is dominated by the Skiddaw Slate Group. This forms the upland terrain peripheral to Bassenthwaite Lake. Radial ice flow from the centre of the Lake District massif throughout the Quaternary has preferentially over-deepened many valleys including Bassenthwaite Lake and Derwent Water which drain the northern Lake District. The Derwent/Bassenthwaite corridor exhibits a range of glacial features that relate to both ice advance and retreat during the Last Glacial Maximum (Hughes *et al*., 2010). Bassenthwaite Lake is situated in the floor of a large, glacially streamlined U- shaped valley. The most prominent glacial geomorphological features peripheral to the lake are roches moutonnees, drumlins, moraines and outwash plains (Figure 5.1). It is not until recently that several studies have demonstrated the existence of submerged glacial features and sediments in the Lake District (Pinson *et al*., 2013) but, as will be demonstrated, several features can be found on the lake floor of Bassenthwaite.

5.1.1. Drumlins

Drumlins were mapped following the criteria outlined by Stokes and Clark (2001), Stokes and Clark (2002), Smith *et al*. (2006) and Spagnolo *et al*. (2010). Concentrated areas of drumlins include the flat, ~ 4 km^2 basin to the south of Bassenthwaite Lake. They also occur to the east of Bassenthwaite Lake. The average long axis of these features is ~ 400 m, with some longer bedforms up to ~ 700 m. Long axes are generally orientated from the south east to the north west. Numerous drumlins were mapped that have not been previously identified (Hughes *et al*., 2010). This is despite difficulties in identifying drumlins owing to postglacial sedimentary infill between drumlins resulting in a more subdued topography.

5.1.2. Moraine ridges

Linear to curvilinear features adjacent to Bassenthwaite and running perpendicular across the valley basin were identified as moraines. Typically they are ~ 300 m in length and ~ 5 – 8 m in height. In a number of localities they clearly run in to the lake (Figure 5.1). These moraines can also be identified from one side of the lake to the other and were observed on the lake floor (Figure 5.9). Moraines are primarily distributed to the north and along Bassenthwaite Lake and are not observed in the area to the south of Bassenthwaite Lake which is dominated by drumlins.

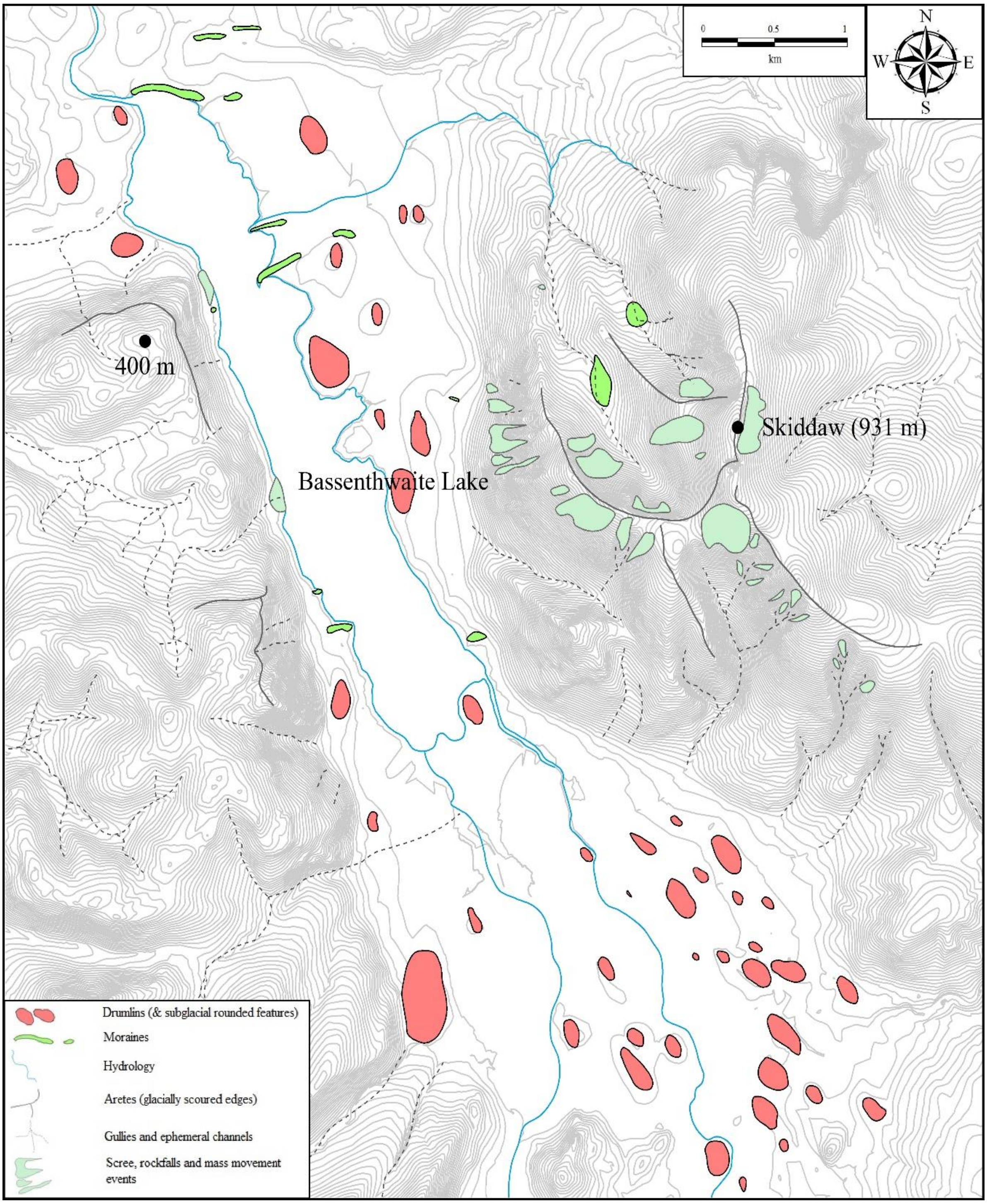


Figure 5.1. The glacial geomorphology of Bassenthwaite Lake and adjacent areas. Palaeo-ice flow would have occurred from the bottom to the top of the map.

5.2. Location of sub-bottom profiles

The location and length of eight sub-bottom transects is shown in Figure 5.2. These eight transects contain detailed information on sub-bottom sedimentary structure and lake-floor glacial geomorphology.

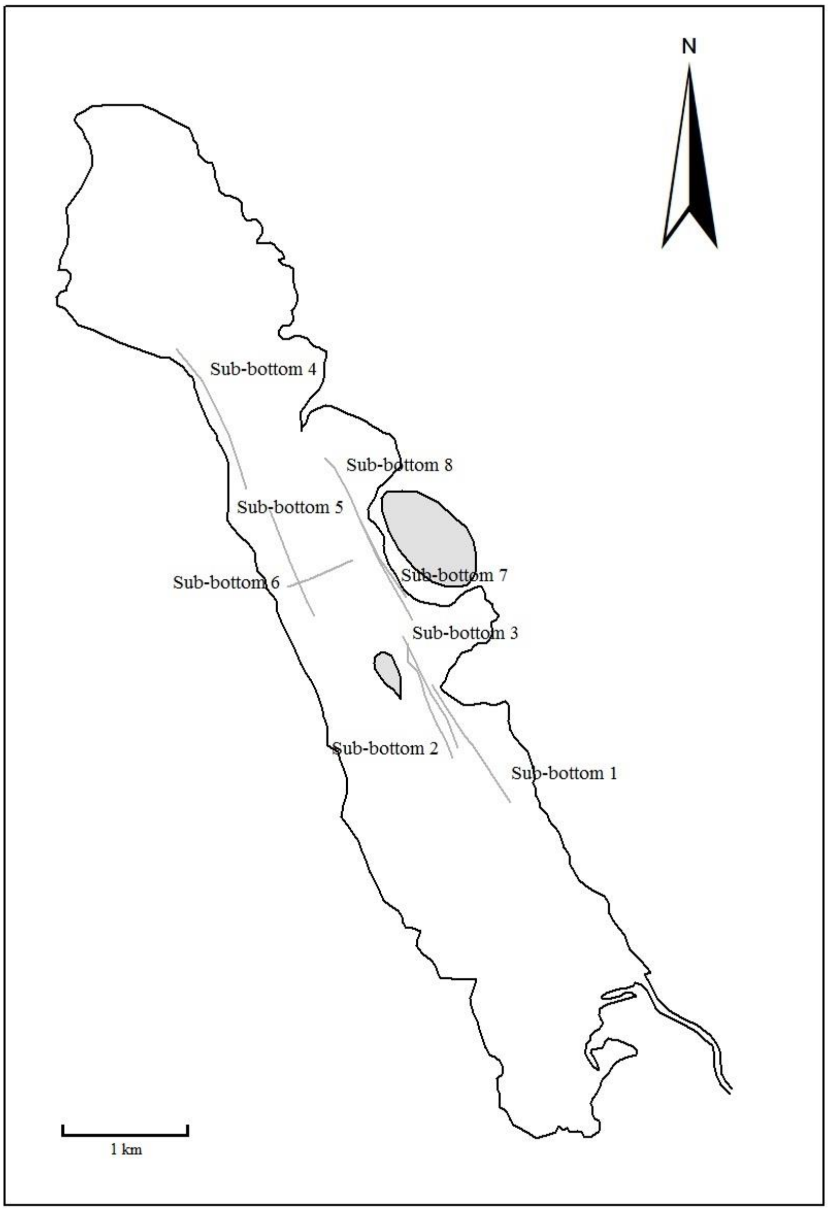


Figure 5.2. The location and length of the 8 sub-bottom transects. The location of a lake marginal and a potential submerged drumlin is also shown.

5.3. Sub-bottom acoustic facies

5.3.1. Facies LF1

LF1 is the lowest unit within the sub-bottom profiles and is observed, to varying extents, within all transects. LF1 is a transparent, homogeneous unit with no internal stratification or deformation. Its lower contact boundary was not observed. The upper contact with LF2 is typified by a sharp transition to a high reflectance, stratified sediment. LF1 has an undulating surface which varies by as much as 5 – 10 m in amplitude and influences the thickness and distribution of above sedimentary facies (Figure 5.3).

5.3.2. Facies LF2

LF2 is the second lowest facies and also occurs within all of the sub-bottom transects. The contact to the underlying bedrock or till facies (LF1) is sharp and the unit has an average thickness of ~ 6 - 8 m. Discontinuous, weak internal stratification is apparent and the unit has high reflectance properties. In places it drapes the underlying bedrock / till (LF1). The boundary between LF2 and the overlying LF3 is generally sharp. LF2 is thickest in areas overlying depressions in LF1, but it can be traced conformably overlying undulations in LF1. The change in thickness and depth of LF2 results in the large changes in lake-floor geomorphology. This facies, although illustrating similar internal properties to the above LF3, differs as it has strong reflections throughout and is thinner than LF3 (Figure 5.3).

5.3.3. Facies LF3

LF3 has an average thickness of ~ 8 m. LF3 occurs within all sub-bottom profiles and overlies LF2 in all transects. It conformably overlies and onlaps LF2. It thickens and thins across the lake floor being thickest in depressions in LF1 and thinning and on-lapping against highs in the lake floor bathymetry. LF3 displays continuous parallel stratification which varies from transparent to high reflectance. This is in contrast to LF2 below which illustrates generally high reflectance throughout. It is sharply overlain in a few localities by LF6 and sharply overlies LF2 in all transects (Figure 5.3).

5.3.4. Facies LF4

LF4 overlies both LF2 and LF3 and is located in Sub-bottom 7 and Sub-bottom 8. This facies conformably overlies LF3 and truncates this unit. The contact to overlying and underlying sedimentary units is sharp. LF4 occurs to a maximum thickness of ~ 2 m and does not change in thickness significantly across sub-bottom transects. LF4 is weakly internally stratified and has strong reflections and the upper and lower bounding surfaces. LF4 illustrates strong reflection properties in comparison to the underlying LF3 unit (Figure 5.3).

5.3.5. Facies LF5

LF5 occurs to a maximum thickness of ~ 3 m. LF5 was observed within 2 sub-bottom transects only (Sub-bottom 6 & 8). Following a gradual increase in thickness within Sub-bottom 8, LF5 reaches its maximum thickness of ~ 4 m at ~ 620 m. Therefore, LF5 occurs as a sedimentary wedge, increasing in thickness from south to north. LF5 occurs between LF6 (above) and LF3 (below) with a sharp contact between these adjacent facies: this lithofacies occurs as a primarily transparent unit, however, faint parallel reflections are apparent in places. This is in contrast to LF1 which illustrates no internal stratification, stronger reflectance and is entirely homogeneous (Figure 5.3).

5.3.6. Facies LF6

LF6 occurs within all sub-bottom profiles and drapes large swathes of the lake-floor. The average depth of this sequence is ~ < 1 m. However, its thickness is dependent upon the lake bathymetry, with thicker accumulations occurring within lake-floor depressions, as illustrated in Sub-bottom 3 (Figure 5.7). Therefore, thinning of this lithofacies occurs on slope locations (Figure 5.3). This acoustic facies represents the youngest depositional sequence and drapes the lake-floor. LF6 is internally stratified and illustrates weak to strong reflections with changes in the stratification (Figure 5.3). LF6 conformably, but sharply, overlies all other lithofacies. No deformation of internal stratification of LF6 is apparent.

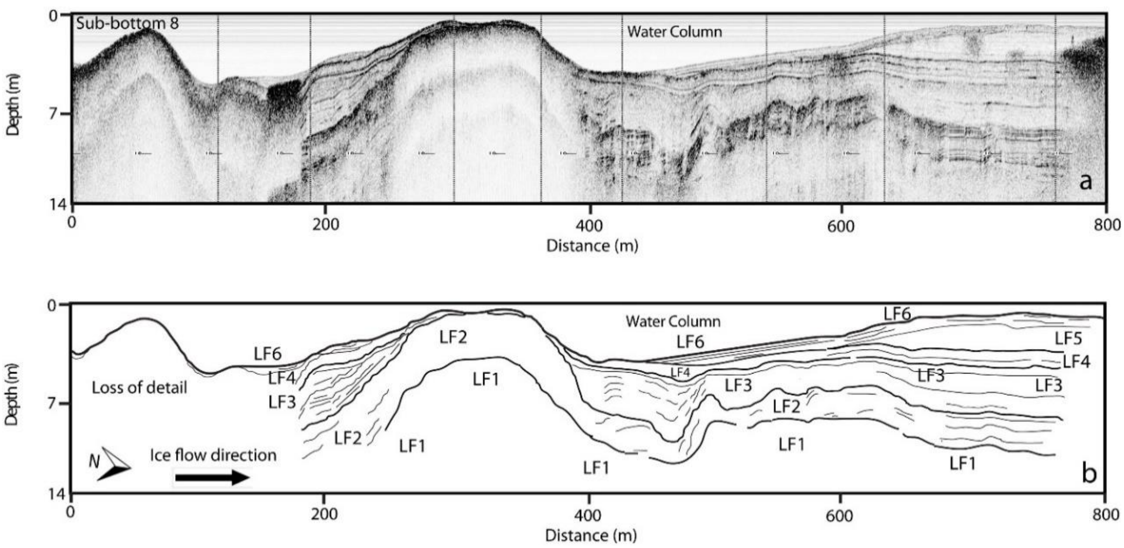


Figure 5.3. Example of all lithofacies described above and their stratigraphic positions within Sub-bottom 8.

5.3.7. Loss of detail

Homogeneous, strong reflections that become increasingly transparent with depth and restrict observations of any potential underlying facies are considered as areas of loss of detail in sub-bottom sampling. A potential explanation for the loss of detail within Bassenthwaite Lake is the presence of shallow gas within the lake sub-surface. This was observed by Pinson *et al*. (2014), within similar sub-surface facies, across Lake Windermere in the south of the Lake District (Figure 5.4). This resulted in large portions of Lake Windermere inaccessible for sub-surface analysis. However, the physical principles which resulted in loss of detail owing to the presence of gas were not established by Pinson *et al*. (2014).

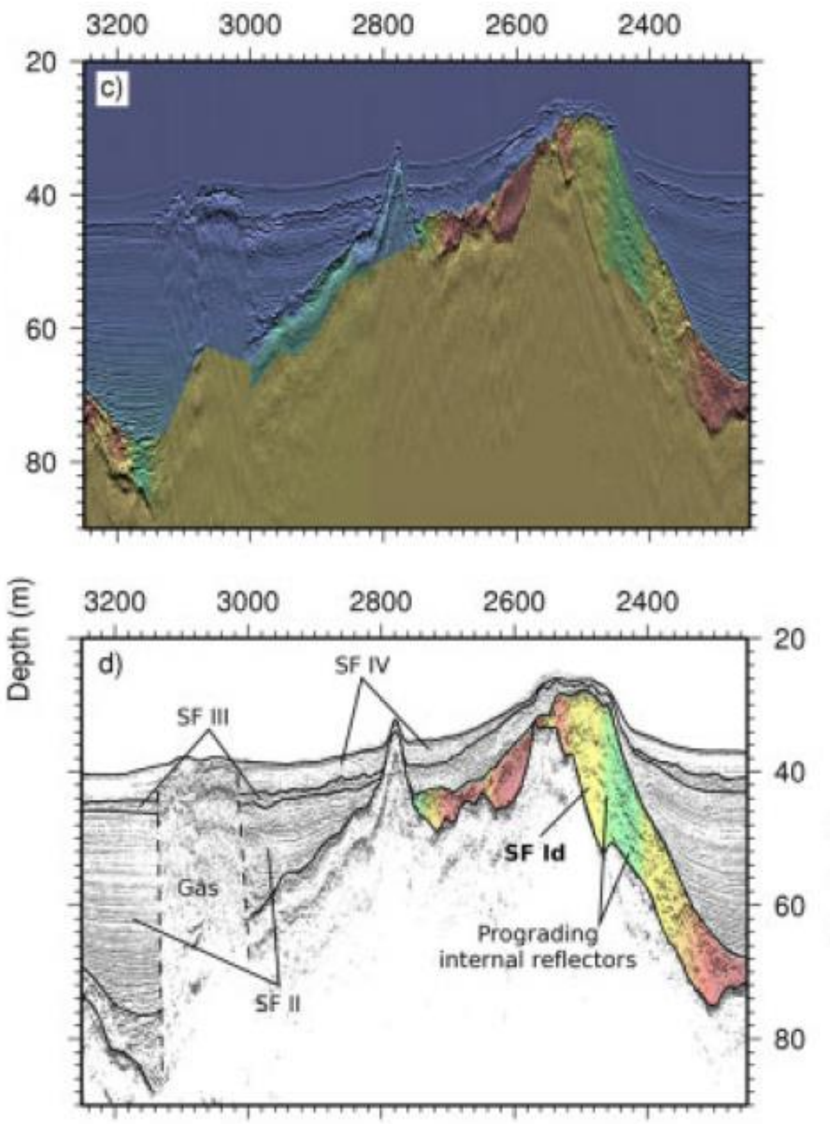


Figure 5.4. Uninterpreted raw data (above) and interpreted image (below) of the sub-bottom composition of Lake Windermere (Pinson *et al*., 2014). Here, a lithofacies of similar acoustic reflectance has been interpreted to result from the presence of gas within the sediment.

5.3.8. Summary of lithofacies

The analysis of 8 sub-bottom transects has resulted in the identification of 6 distinct acoustic facies (Figure 5.5).

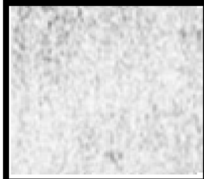
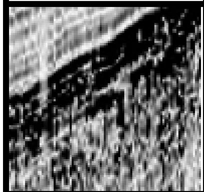
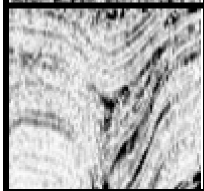
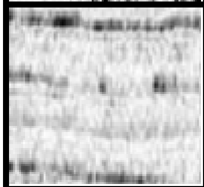
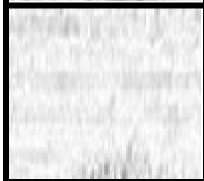
Acoustic Facies (CHIRP)	Label	Description
	LF1	Homogeneous facies with no internal stratification. Illustrates the lowest recorded facies. No internal deformation apparent.
	LF2	Weak internal stratification which occurs discontinuously. Strong reflections observed throughout the facies. Illustrates the second lowest recorded facies. Some deformation.
	LF3	Continuous parallel reflected facies with a weak to strong reflected stratification occurring. Deformation apparent throughout facies. Occurs to a maximum thickness of > 10 m and conformably overlies and onlaps LF2 below
	LF4	Partial internal stratification of high reflectance. No deformation apparent. Occurs to a maximum thickness of ~ 2 m. Conformably overlies LF3 below.
	LF5	Transparent, very weak lateral stratification. Occurs to a maximum thickness of ~ 4 m. Differs from other facies owing to no internal deformation and very weak internal stratification.
	LF6	Parallel reflected, continuous, internally stratified facies overlapping all other facies. Weak to strong reflected stratification occurring at cm-scale. A lack of internal deformation is apparent. This facies occurs to a maximum thickness of ~ 2 m

Figure 5.5. Summary of the main sub-surface facies with an example image, label allocation and brief description provided.

5.4. Sub-bottom Profiles – Lithofacies Architecture

5.4.1. Sub-bottom Profile 1

Sub-bottom 1 (Figure 5.6) is a 500 m long transect, obtained towards the south of Bassenthwaite Lake (Figure 5.2). LF1 has an undulating surface and thus influences the thickness and distribution of overlying facies. LF2 conformably onlaps LF1 and often thickens in depressions as it offlaps high amplitude undulations in LF1. LF3 onlaps LF2 and occurs primarily within depressions of LF1 (for example, at ~ 100 m and 250 m (Figure 5.6)). Unfortunately, lack of detail prevents large-scale descriptions of sub-surface stratigraphy within this transect. LF6 drapes all other facies and a thick accumulation of this facies occurs at ~ 375 m onwards. LF6 occurs very thinly (if at all) on top of the high amplitude LF1 undulations.

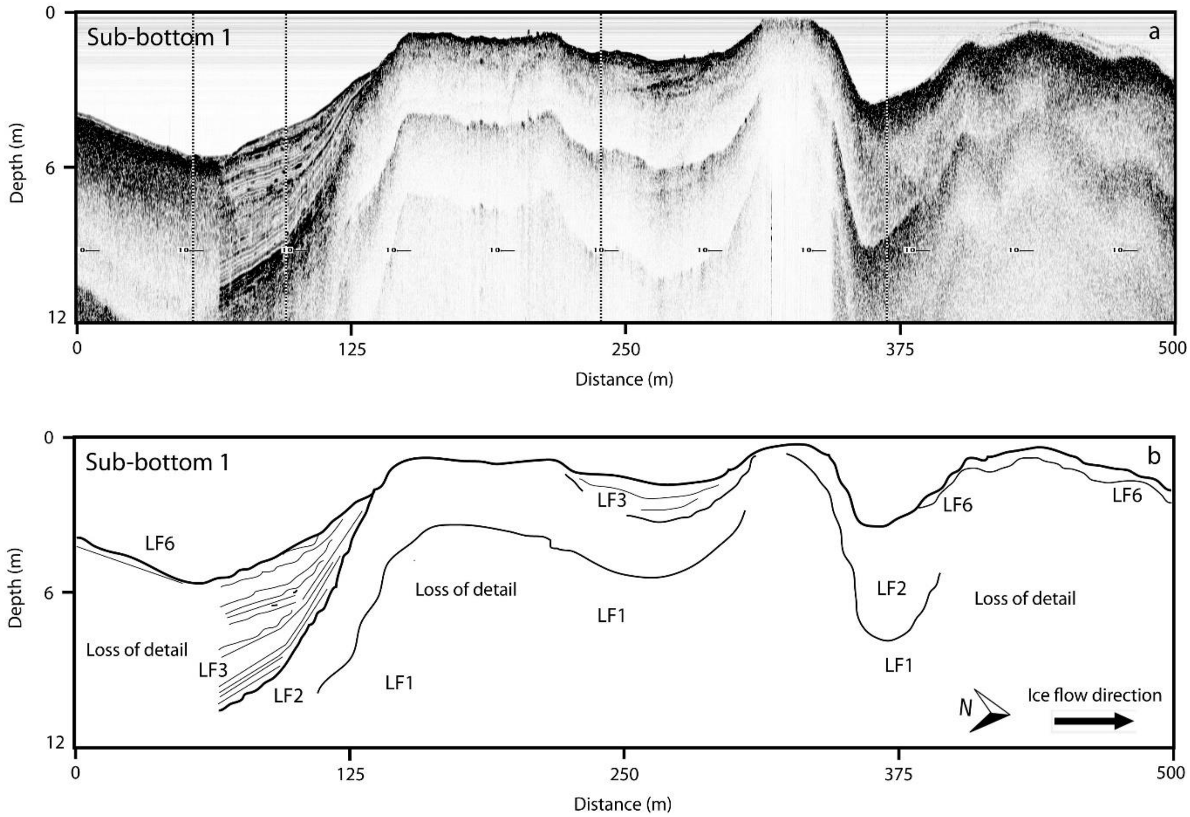


Figure 5.6. A 500 m lateral sub-bottom profile with loss of detail occurring and with LF2, LF3 and LF6 also apparent.

5.4.2. Sub-bottom Profile 2

Sub-bottom 2 (Figure 5.7) was also obtained from a southern location within Bassenthwaite Lake (Figure 5.2), adjacent to Sub-bottom 1. Despite a loss of signal in many sectors of this transect it is possible to see LF1 undulations as observed within Sub-bottom 1. Both LF2 and LF3 occur in thick sequences within depressions in LF1 and occur as conformable, overlying stratified sequences. Thicker accumulations of LF2 and LF3 include areas at ~ 500 and 650 m. LF6 primarily occurs either side of the ~ 10 m high undulation, with sloping areas resulting in thicker accumulations of this facies. An infilling of LF2 by LF3 occurs at ~ 550 m within this transect and results from a depression within the surface of the LF2 below.

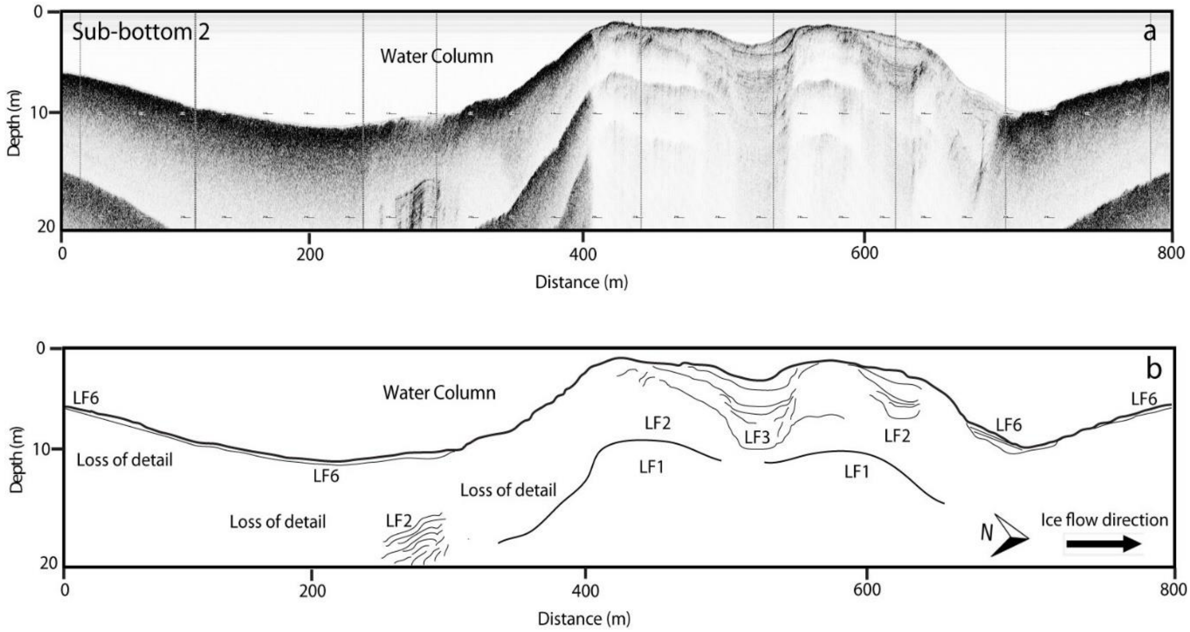


Figure 5.7. Sub-bottom 2, 800 m in length and 20 m in depth. A relatively small, ~ 30 m detailed area occurs within this profile beneath a loss of detail area. Palaeo-ice flow would have occurred from left to right of the image.

5.4.3. Sub-bottom Profile 3

Sub-bottom 3 (Figure 5.8) is the final transect taken from the southern end of Bassenthwaite Lake (see Figure 5.2). Clear undulations in are conformably overlain and draped by LF2. This is followed by LF3 which onlaps and overlies LF2. LF3 also occurs conformably above LF2 with a sharp transition separating these facies. LF3 occurs to a thickness of ~ 7 m and differs from LF2 with stratification illustrating strong to weak reflectance. Variations in the thickness of internal stratification occurs within LF3, with a generalised thinning of stratification occurring towards the top of LF3. LF6 thins on lake-floor slopes and accumulates in flat areas. Within LF6, two submerged features occur at ~ 450 m and 680 m and appear at the interface between LF3 and LF6.

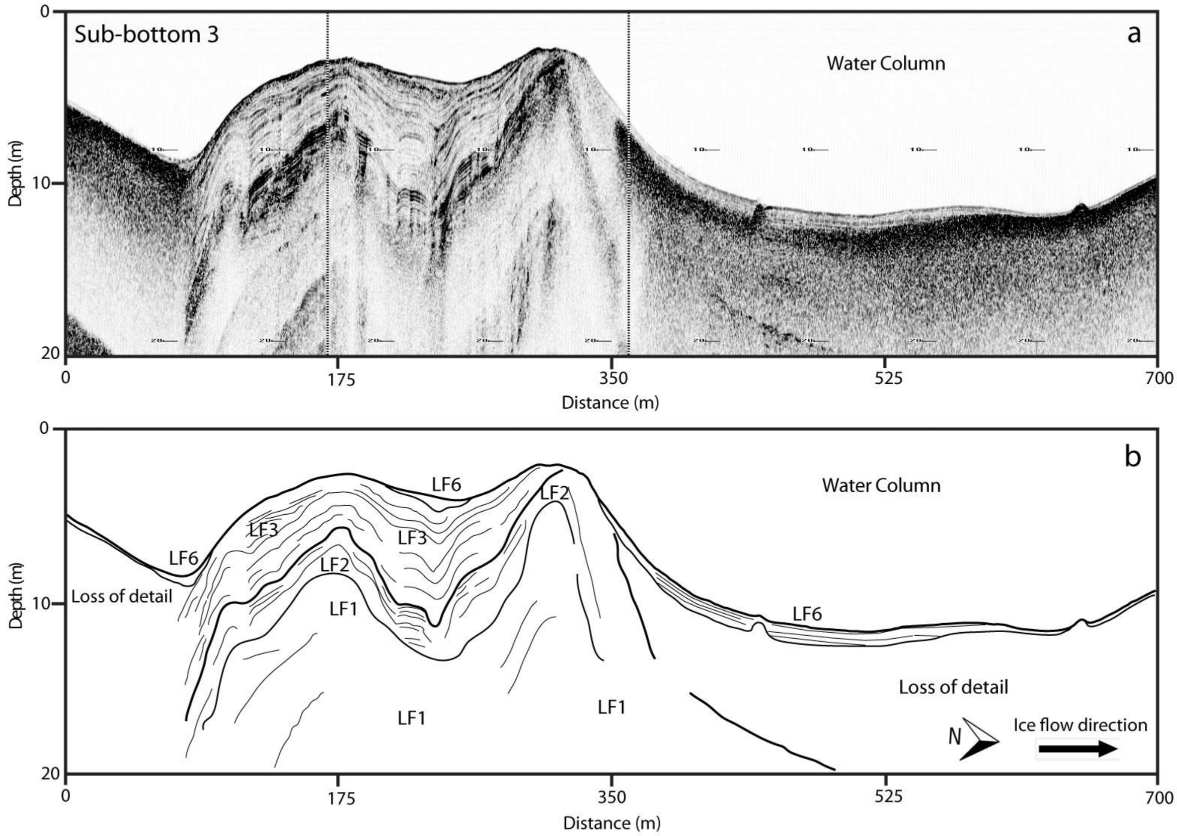


Figure 5.8. Sub-bottom profile transect 3 which recorded the undulations in LF1 and subsequent overlying sedimentation. Palaeo-ice flow would have occurred from left to right of the image.

5.4.4. Sub-bottom Profile 4

This transect (Figure 5.9) is the most northern subsurface area considered within Bassenthwaite Lake (Figure 5.2) and includes geomorphological information. A small ~ 30 m long and 5 m high undulation within Sub-bottom 4 is composed of LF1 and LF2 and onlapped by LF3. Is not draped by LF6. LF2 again occurs as an overlying, continuous facies above LF1 with LF3 onlapping these facies. A change in sediment thickness occurs either side of this undulation. LF6 occurs here to its greatest thickness within all transects at ~ 2 m with internal stratification clear and continuous throughout the facies.

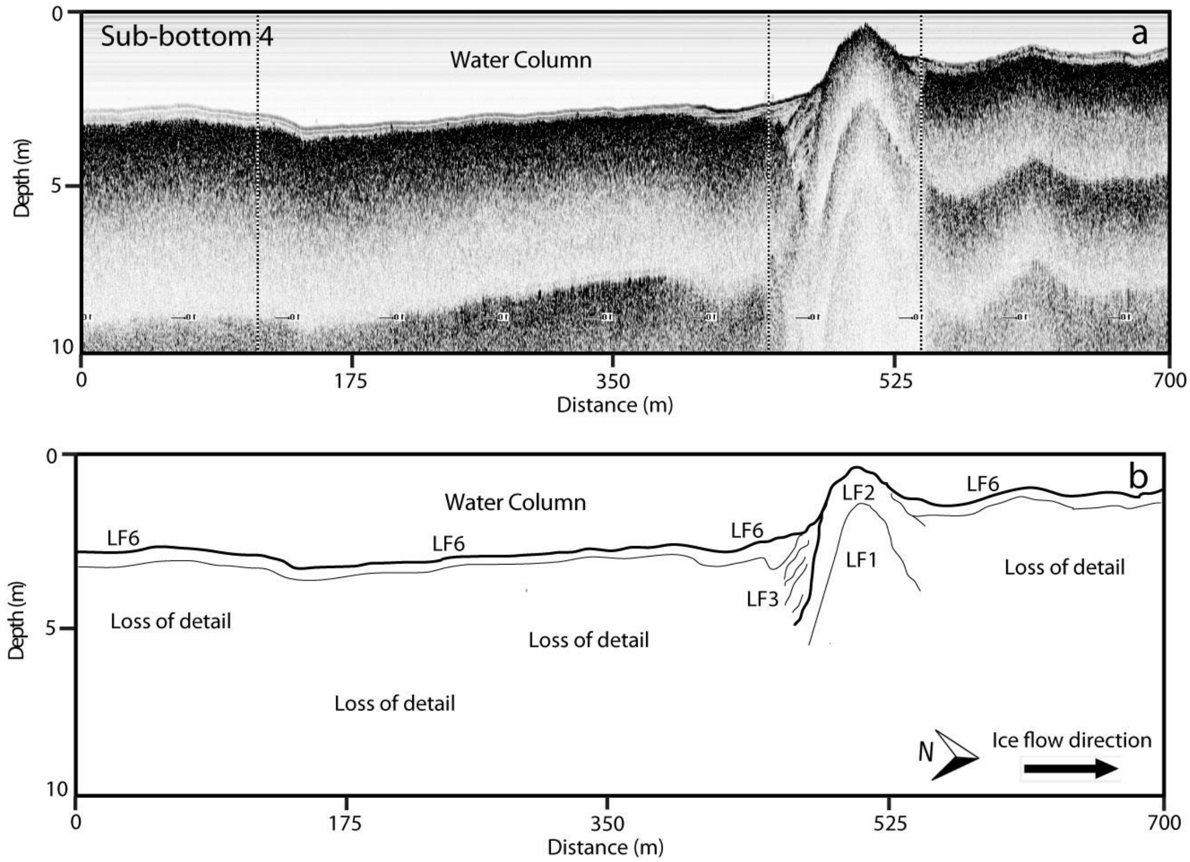


Figure 5.9. Sub-bottom 4 profile, with palaeo-ice flow occurring from left to right of the image, illustrating a generalised horizontal lake floor, with a lack of recorded detail. This is punctuated by an undulation in LF1, followed by LF2 sedimentation.

5.4.5. Sub-bottom Profile 5

Sub-bottom 5 (Figure 5.10) was collected towards the centre of Bassenthwaite Lake (Figure 5.2). This transect was also obtained from the deepest area of the lake at ~ 20 m below the lake surface. LF1 is apparent from 100 to 200 m, with the surface of LF1 decreasing in height and resulting in a thicker accumulation of LF2 and LF3. Here, LF2 and LF3 exhibit vertically stacked, conformable on-lapping strata of variable grain size. Here, the higher reflectance of LF2 in comparison to LF3 is illustrated, despite the similar internal stratification characteristics. A relatively subdued lake-floor geomorphology is apparent within this transect resulting from the lack of large undulations in LF1.

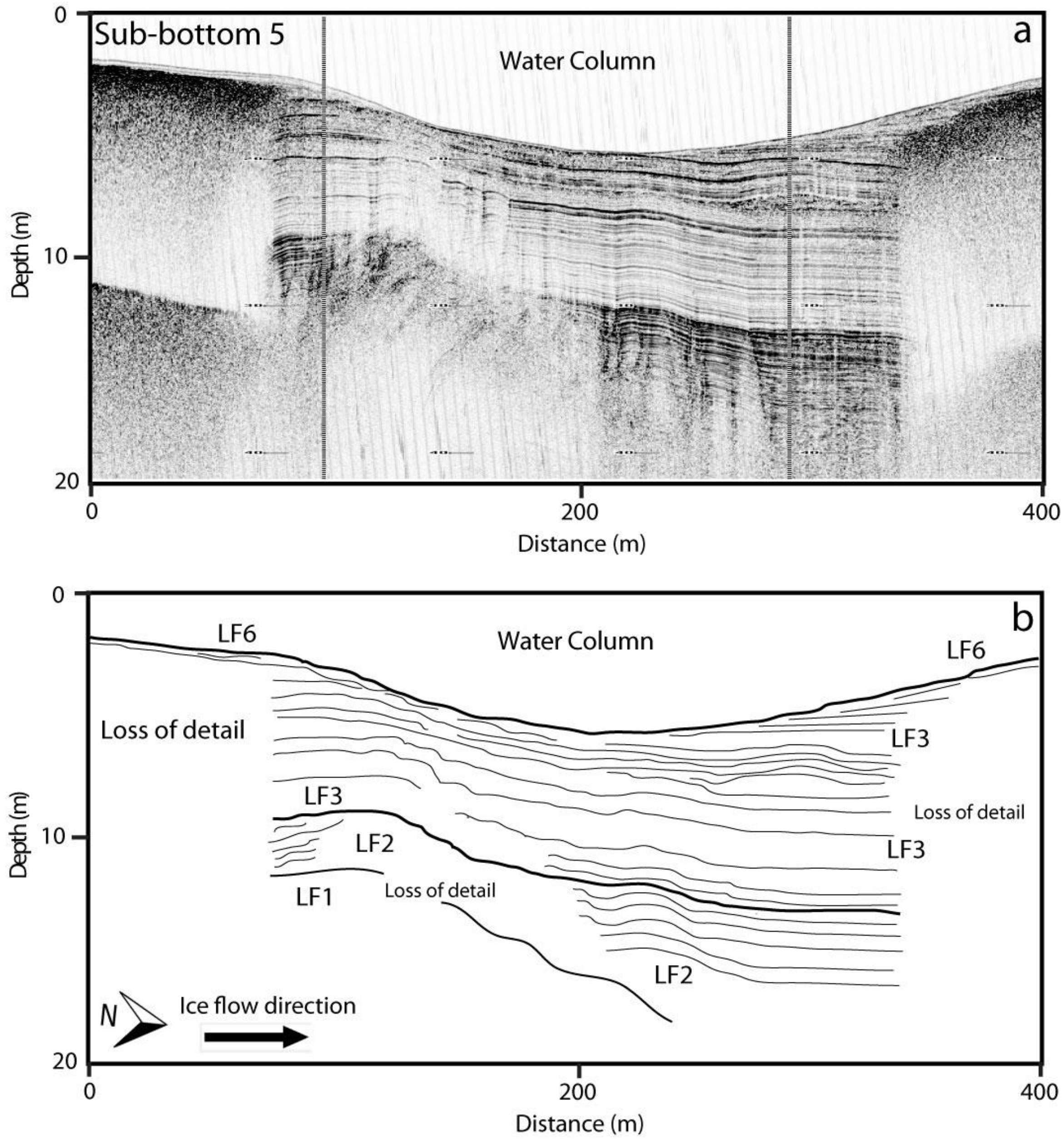


Figure 5.10. The relatively subdued geomorphology of Sub-bottom 5 with LF1, LF2, LF3 and LF6 apparent. Palaeo-ice flow would have occurred from left to right of the image.

5.4.6. Sub-bottom Profile 6

Sub-bottom 6 (Figure 5.11) was collected east to west, across Bassenthwaite Lake (Figure 5.2) and intersects Sub-bottom 5. This 600 m transect contains LF1, LF2, LF3, LF5 and LF6. Here, as in Sub-bottom 5, loss of detail occurs at the start and end of the transect. A small undulation of LF1 is also observed here. The decrease in height of the undulation results in thicker sediment accumulations of LF2 and LF3. The undulation of LF1 controls the in the change in lake-floor topography. Again, the difference in internal reflection between LF2 and LF3 is apparent, despite similarities in internal composition and structure. LF5 also occurs within this transect and appears to onlap LF3 with a sharp boundary and a change in bedding direction separating these.

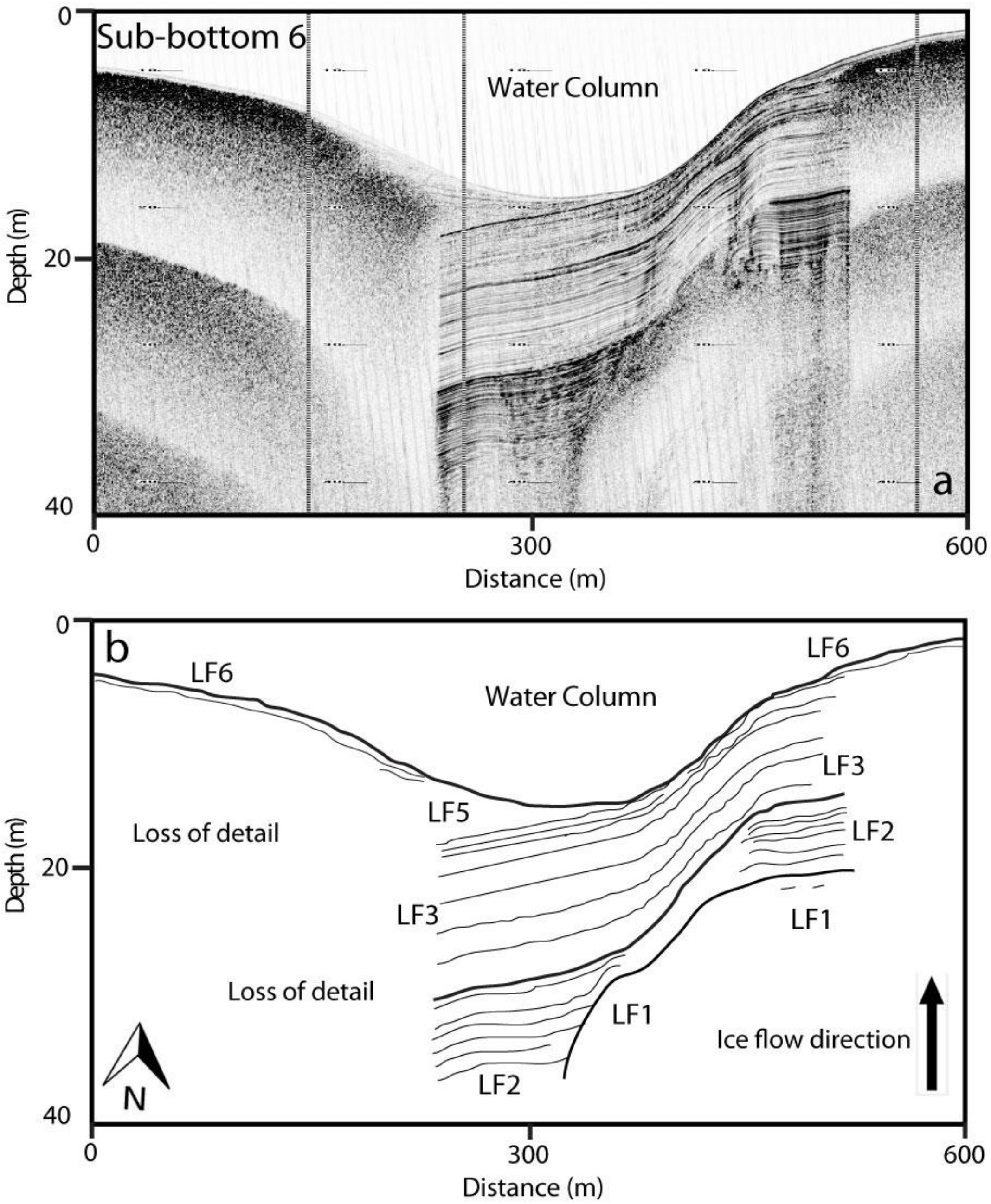


Figure 5.11. A 600 m long, 20 m deep sub-bottom profile illustrating all acoustic faces. Here, palaeo-ice flow would have occurred perpendicular to the image. Note the similarity between lake-floor topography and the undulation within LF1.

5.4.7. Sub-bottom Profile 7

Sub-bottom 7 (Figure 5.12; 5.14) is one of two transects taken adjacent to a drumlin that occurs outside of the lake that was previously identified by Hughes *et al*. (2010) (Figure 5.2). Sub-bottom 7 and Sub-bottom 8 are adjacent transects (within ~ 20 m) and provide a lateral and longitudinal consideration of subsurface stratigraphy. Within Sub-bottom 7 LFs 1, 2, 3, 4 and 5 can be identified. Here, the lake-floor topography and subsequent stratigraphy is influenced by the undulation in LF1. This undulation occurs to a height of ~ 10 m and length of ~ 250 m. LF2 then conformably drapes LF1 and occurs to an average thickness of ~ 3 m. There is a sharp contact between the homogeneous LF1 and the stratified LF2. LF3 occurs to a thickness of ~ 4 m and again conformably overlies LF1 and LF2. The thickness of LF3 is influenced by the underlying geometry of LF1, with thicker accumulations occurring within depressions of LF1. LF6 drapes these facies.

5.4.8. Sub-bottom Profile 8

Sub-bottom 8 (Figure 5.13; 5.14) was obtained adjacent to Sub-bottom 7 (Figure 5.2). This transect is the longest transect with detailed stratigraphy with LF1, LF2, LF3, LF4, LF5 and LF6 occurring. The first 400 m illustrates lateral changes between transects and again illustrates an undulation in LF1, followed by the overlying, conformable sequences of LF2 and LF3. LF2 occurs continuously across this transect and does not vary substantially in thickness. The internal stratification within LF2 is not as obvious as that observed in LF3 with high reflectance throughout. Areas of no stratification are also apparent within LF2. Following this LF3 infills depressions within the underlying LF1. Variations in the thickness of laminations occurs throughout LF3 and changes in reflectance also occurs. LF5 occurs within this transect only and is a transparent facies which increases in thickness along the transect. LF6 occurs above all facies, as it does at all locations.

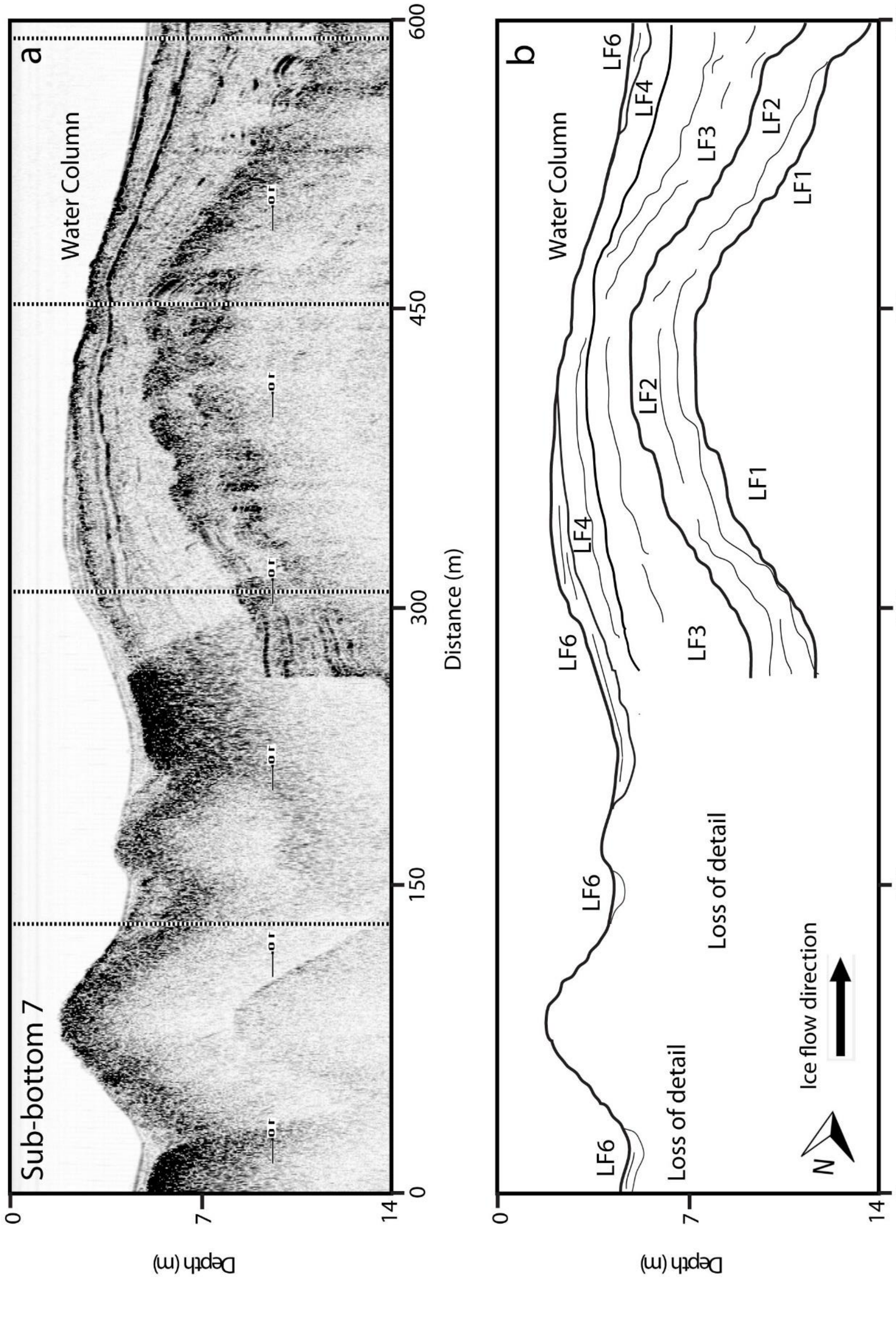


Figure 5.12. Sub-surface composition and structure of a drumlin. Here, ice would have flowed from left to right of the image.

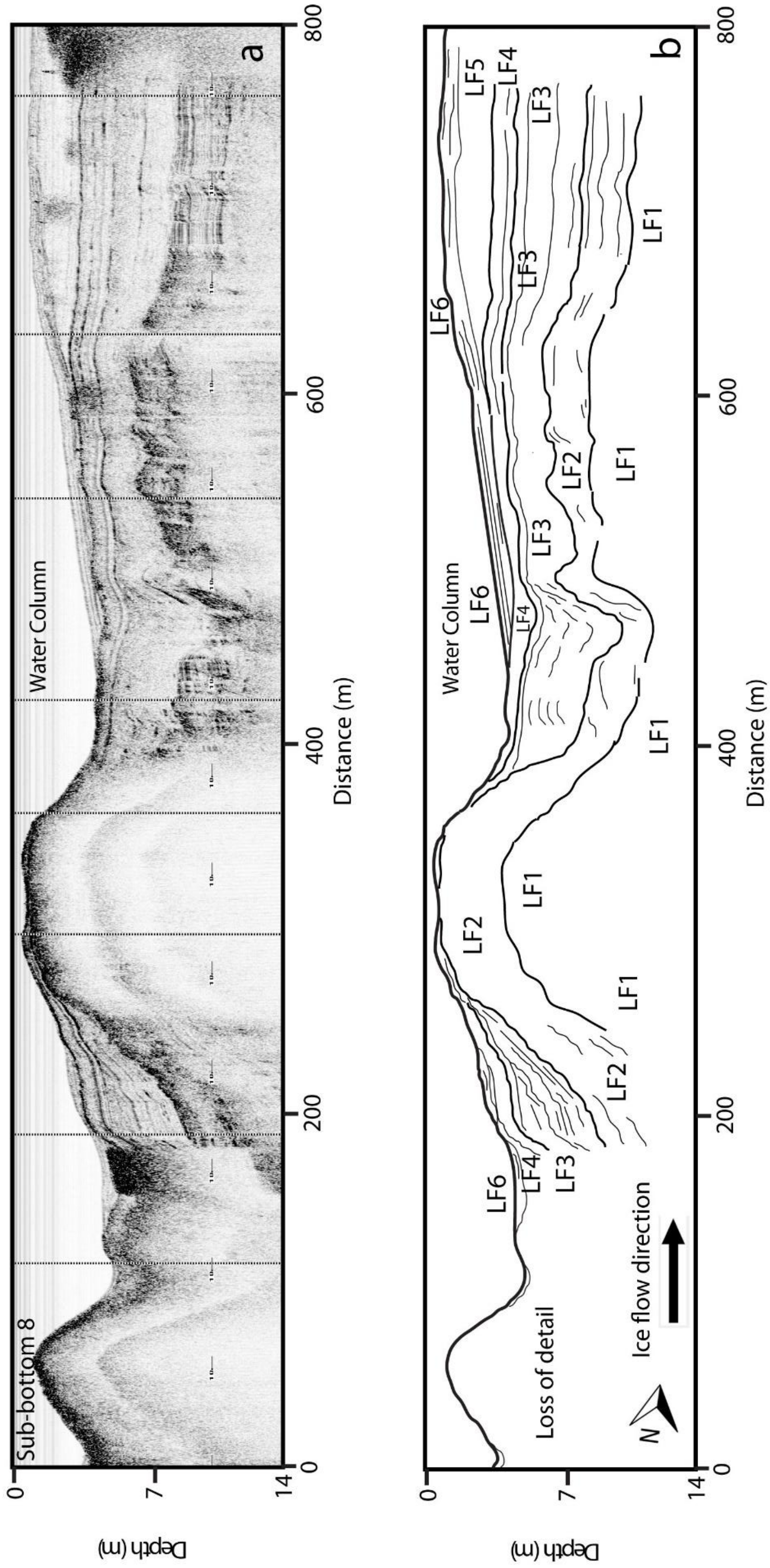


Figure 5.13. The longest transect of detailed, sub-surface information is provided here and includes the relatively sparse LF5, occurring as a wedge infill and overlying all lithofacies except from the youngest, drape facies (LF6). Here, as above, ice would have flowed from right to left of the transect.

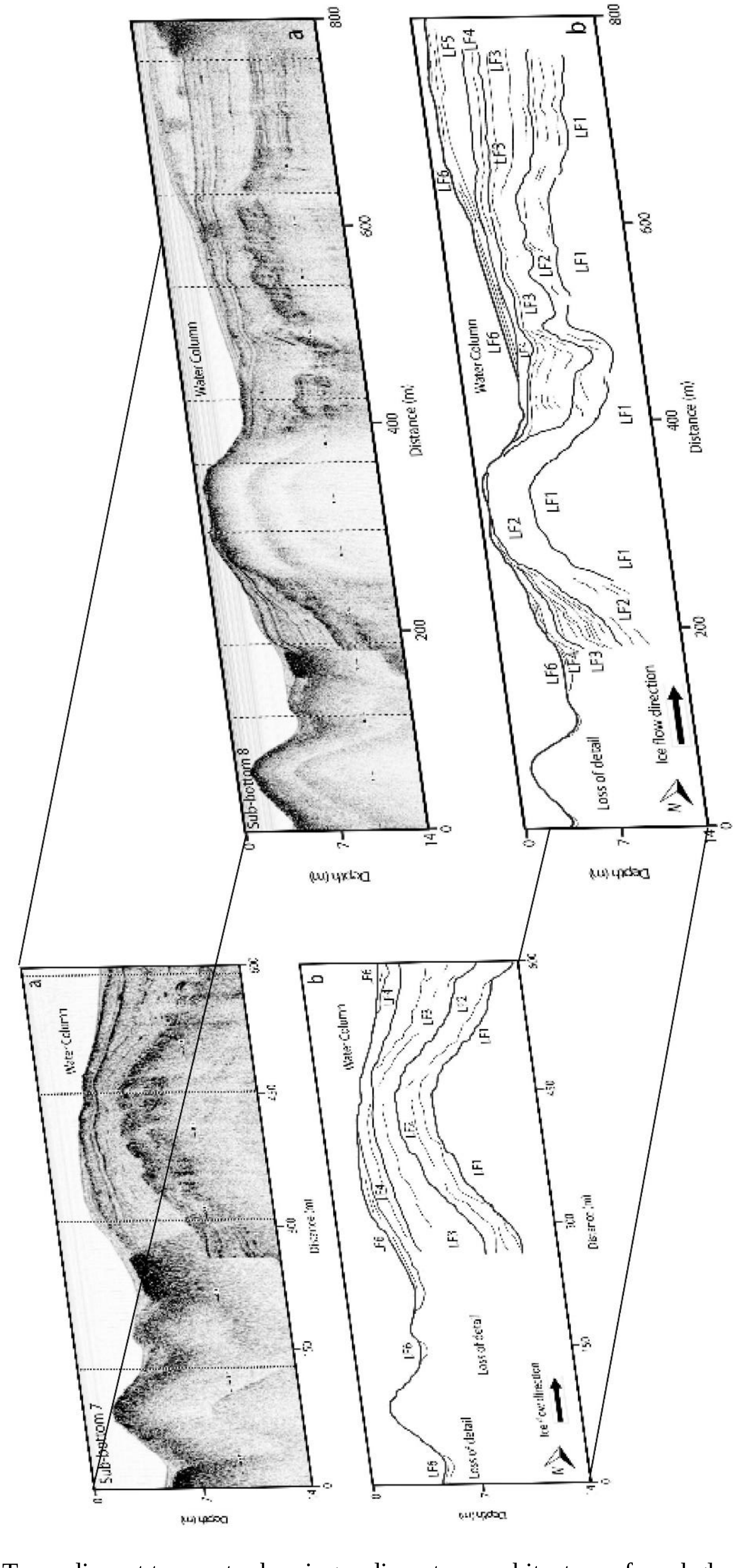


Figure 5.14. Two adjacent transects showing sedimentary architecture of a subglacial, lake-floor drumlin and illustrating lateral changes in sedimentary architecture.

6. INTERPRETATION

6.1. The glacial geomorphology of Bassenthwaite Lake and adjacent areas

The geomorphological map of Bassenthwaite Lake shows a suite of glacial landforms that relate to the Last Glacial Maximum (LGM) and subsequent deglaciation (Figure 5.1). Drumlins are the most prominent feature in the vicinity of the lake and converge towards the southern end of the Bassenthwaite basin, illustrating that former ice flow was channelled through the valley towards the north. The drumlins and sediments associated with them represent the oldest phase of glacial landform construction in the valley. The deposition of moraines throughout the Bassenthwaite basin and to the north illustrates final deglaciation (as prior advance would have deformed and overrun these features) with numerous ice-marginal ridges suggesting punctuated, gradual retreat of the ice. The glaciofluvial and glaciolacustrine sequences that form the lake floor relate to a final deglacial phase deposition. Since the final phase of deglaciation, fluvial, subaerial and slope processes have resulted in the formation of scree slope formations and fluvial incision of the valley floor and sides.

The channelling of ice through the Bassenthwaite basin is also reported by Hughes *et al*. (2010) and Livingstone *et al*. (2008) with a convergence of subglacial bedforms in the southern end of the basin apparent. To the north of Bassenthwaite lake Livingstone *et al*. (2008) and Hughes *et al*. (2010) illustrate that the long axis of these subglacial bedforms migrate towards the west and thus show ice draining into the Solway lowlands and northern Irish Sea Basin. The Bassenthwaite basin therefore operates as a major outlet channel for ice radiating from the Lake District ice cap and records numerous phases of ice advance and retreat.

6.2. Interpretation of sub-bottom acoustic facies in Bassenthwaite Lake

6.2.1. Summary of lithofacies interpretations

Five distinct lithofacies were identified from the sub-bottom profiles taken within Bassenthwaite Lake. The internal characteristics of each lithofacies and an interpretation of their deposition are summarised within Figure 6.1 and their stratigraphic relations are illustrated in Figure 6.2.

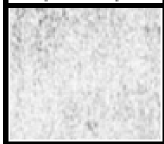
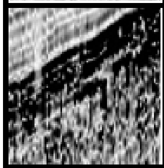
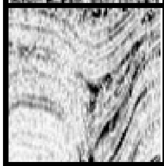
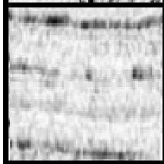
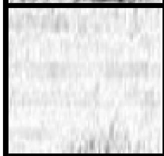
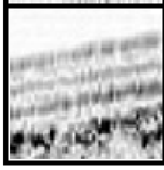
Acoustic Facies (CHIRP)	Label	Interpretation
	LF1	Bedrock / Till
	LF2	Lower glaciolacustrine sequence
	LF3	Upper glaciolacustrine / lacustrine sequence
	LF4	Re-advance till unit
	LF5	Glaciofluvial outwash
	LF6	Holocene lake infill

Figure 6.1. Example images and interpretations of the main sub-surface facies.

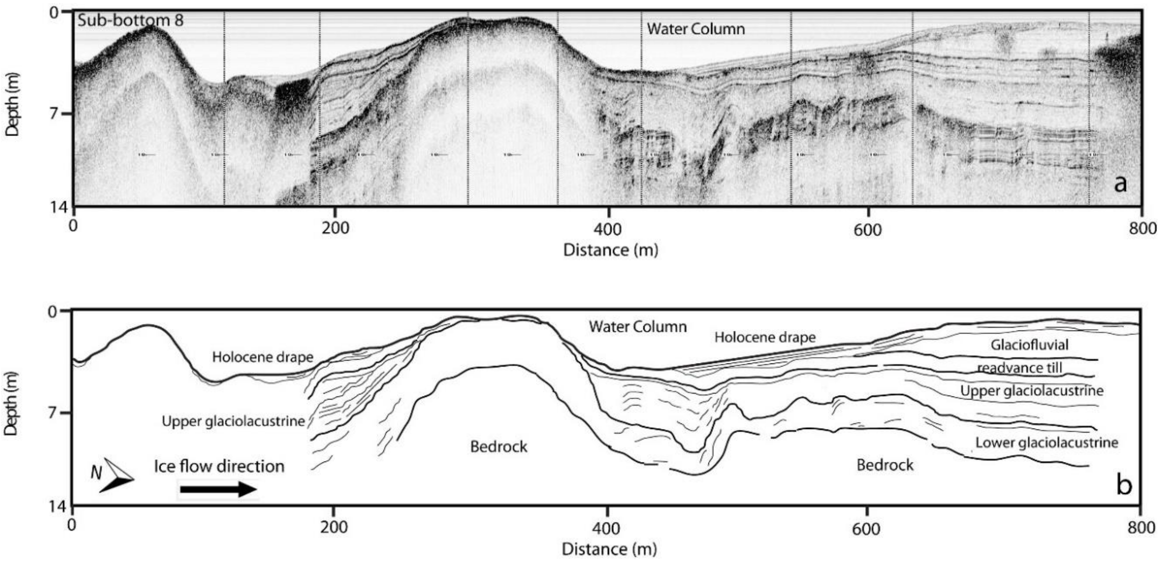


Figure 6.2. Stratigraphic relation of the interpreted lithofacies within the sub-surface of Bassenthwaite Lake.

6.2.2. Facies LF1: Bedrock / Till

This transparent, entirely homogeneous lowest facies is interpreted as either bedrock (Skiddaw Slate Group) or till. It has homogeneous characteristics both vertically and laterally. It occurs to > 20 m in depth and no bounding surfaces are observed below this. Furthermore, undulations on the surface of this facies does not alter the internal characteristics or structure (which is completely transparent and homogeneous). The rounded, undulating surface of LF1 is interpreted here to illustrate potential glacial abrasion and scour of the bedrock resulting from ice override. This interpretation is supported by bedrock outcrops exposed either side of Bassenthwaite Lake which are clearly abraded and scoured (Figure 3.3). A possible till facies is considered because, in some locations, high-angle ridges of LF1 occur. This bedrock or till facies is draped, infilled and onlapped by overlying sediments and forms the undulations on the lake floor. The bedrock / till unit has the significant influence on the geomorphology of the lake-floor.

A similar, transparent and homogeneous lowest facies was also identified by Pinson *et al*. (2013) within Lake Windermere and was interpreted as the acoustic basement. Pinson *et al*. (2013) also identified subsequent glaciolacustrine and glaciofluvial sedimentation infilling such bedrock undulations. The lowest facies identified by Pinson et al. (2013) also exhibits an undulating, sharp contact surface to the overlying sediments. A rounded, eroded and undulating bedrock basement was also identified by Waldmann *et al*. (2010) in Lago Fagnano, southern Patagonia. Waldmann *et al*. (2010) obtained seismic profiles within Lago Fagnano and identified a transparent and homogeneous basement unit. The rounded, eroded surface of this basement unit was also interpreted, as is here, to result from previous glacier overriding and scouring of the bedrock.

6.2.3. Facies LF2: Lower glaciolacustrine

This internally stratified, strongly reflected lithofacies is interpreted here as a lower glaciolacustrine facies. Both this facies and the overlying glaciolacustrine facies (LF3) form the bulk of deposits within Bassenthwaite Lake. This facies is internally stratified, suggesting alternation is grain size and internal strata characteristics. The alteration in strata may result from changes in meltwater or sediment flux or location of the drainage source adjacent to the retreating ice mass. This lithofacies, in most locations, drapes the underlying bedrock / till facies and is significantly thinner than the overlying lithofacies. As LF2 continuously drapes the lake-floor it denotes rapid, perhaps synchronous deposition of glaciolacustrine sediment across the lake basin as the ice retreated. As there is no wide-spread deformation apparent within this facies it appears that no further, large-scale glacier overriding of the basin took place.

The interpretation of a glaciolacustrine lithofacies is consistent with interpretations of a similar lithofacies associated with ice retreat in the Lake District (Pinson *et al*., 2013; Figure 6.3). In Windermere similar lithofacies illustrating stratification and changes in reflectance were associated with changes in grain size. Furthermore, the sediment clearly draped the underlying till facies and formed part of a two-tier glaciolacustrine sequence (as is apparent within Bassenthwaite Lake). Pinson *et al*. (2013) then correlated the acoustic facies from Windermere to sediment cores taken from the lake bottom and found they correlated to a laminated pink clay. A similar, thinly bedded, drape glaciolacustrine facies was also observed within deglacial sediments in a Swiss alpine Lake (Fiore *et al*., 2011). Fiore *et al*. (2011) associated this facies with ice-proximal deposition.

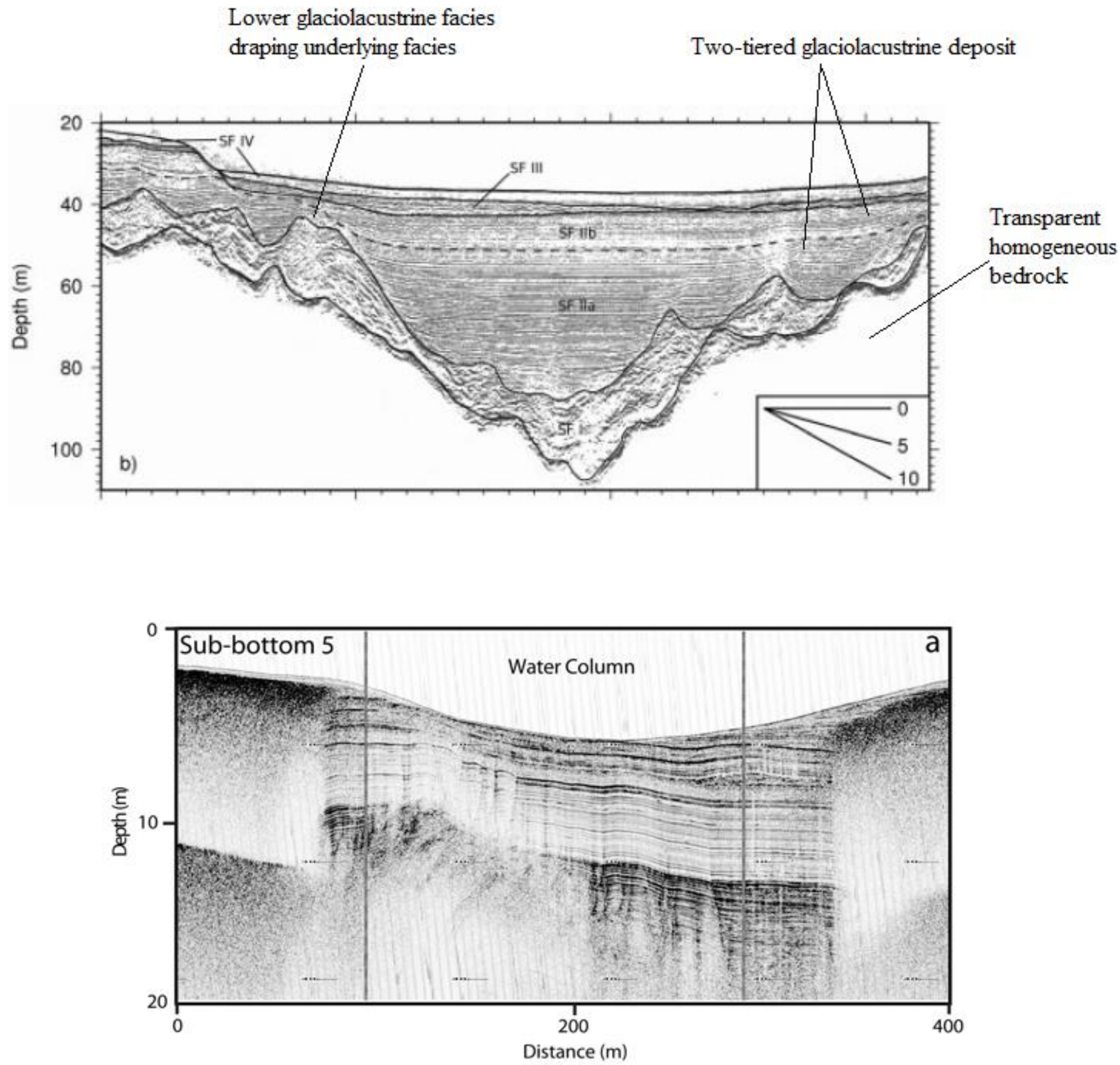


Figure 6.3. Example of similar lithofacies observed by Pinson *et al*. (2013) within Lake Windermere (above). These include a transparent, homogeneous acoustic basements and a two-tiered glaciolacustrine deposit. A transect illustrating two glaciolacustrine facies overlying an acoustic basement is also illustrated from within the sub-surface of Bassenthwaite Lake (below).

6.2.4. Facies LF3: Upper glaciolacustrine

This continuously parallel stratified facies which varies from transparent to high reflectance is interpreted as an upper glaciolacustrine facies. The changes from transparent to high reflectance illustrate changes in grain size and reflect, as the lower glaciolacustrine facies does, changes in meltwater and sediment flux or location of drainage areas from a retreating ice mass (a proximal to distal transition). A lack of deformation within this illustrates deposition in a lake basin during deglaciation. Changes in thickness of the reflected layers are interpreted to illustrate changes in sediment accumulation with alterations in sediment input. This glaciolacustrine deposit infills and overlies the lowest bedrock / till unit and subdues the topography of the lake floor. LF3 differs from LF2 in that it tends to infill areas between bedrock / till undulations as opposed to draping the entire lake floor. Further to this, the internal stratification varies from transparent to high reflectance as in contrast to LF2. This is interpreted to reflect the increasingly distal nature of the sediment as the ice retreated southwards. This is supported by dropstone features within LF3 and ice-berg grounding features (Sub-bottom 5; Section 6.3.10).

Two adjacent glaciolacustrine deposits, with different internal characteristics, have been observed in other locations, including glaciolacustrine sedimentation into a Swiss alpine Lake (Fiore *et al*., 2011). In addition to this, two distinct adjacent glaciolacustrine facies were observed by Pinson *et al*. (2013). In both cases, it was interpreted that such changes in internal characteristics illustrate a change in position of the glacier, with a lower glaciolacustrine deposit resulting from a proximal ice margin deposition into a large water body. The topmost facies consists of rhythmically banded fine-grained mud and is interpreted to have been deposited in a more distal environment.

6.2.5. Facies LF4: Re-advance till

This relatively thin, partially internally stratified facies with areas of high reflectance is interpreted as a re-advance till unit. This unit is interpreted as a re-advancing till unit due to it truncating the underlying glaciolacustrine sediments. The interpretation of a re-advancing till unit is also supported by the subsequent deposition above of a glaciofluvial sequence. This transition from fine glaciolacustrine deposits to a coarser glaciofluvial deposition requires either a more proximal ice mass or larger meltwater input. It is interpreted that the presence of this re-advance till unit is relatively localised as it is present within Sub-bottom 7 and Sub-bottom 8 only. The re-advance till unit remains laterally consistent across transects and this is in contrast to the infilling and onlapping glaciolacustrine units. The creation of a drumlin through multiple, laterally consistent till units is also provided by Johnson *et al*. (2010). A high reflectance boundary between underlying glaciolacustrine sediments and the re-advance till unit illustrates a clear alteration in depositional phases. A weakly stratified, relatively thin unit was also identified by Pinson *et al*. (2013), lying above all glaciolacustrine sediments, however in this case it was interpreted as a mass-movement deposit.

6.2.6. Facies LF5: Glaciofluvial outwash

The transparent, homogeneous acoustic facies with very weak internal stratification is interpreted as a glaciofluvial outwash deposit. A late phase of deposition is envisaged for this unit (possibly Lateglacial) and a lack of deformation suggests no later phases of glacier overriding. The geometry of this facies suggests it is a prograding wedge or fan of glaciofluvial material (Figure 5.13; 5.14) with termination of deposition possibly resulting from a change in channel location, a drop in discharge or further ice margin retreat. This facies only occurs in Sub-bottom 6 and Sub-bottom 8 (Figure 5.11; 5.13) and is overlain by LF6. The sharp contact to both overlying and underlying lithofacies and a lack of incision by LF6 into LF3 may suggest rapid deposition of the fan, with a sedimentary 'pulse' of glaciofluvial material at the ice margin.

A glaciofluvial facies is also interpreted owing to similar acoustic reflection properties of a glaciofluvial sequence observed in the sub-surface of Loch Ness (Turner *et al*., 2012). Examples of such outwash facies are illustrated in Figure 6.4. An example of glaciofluvial sedimentation occurring over glaciolacustrine deposits during ice retreat is also illustrated in Figure 6.5 and was observed adjacent to Lake Pukaki, New Zealand.

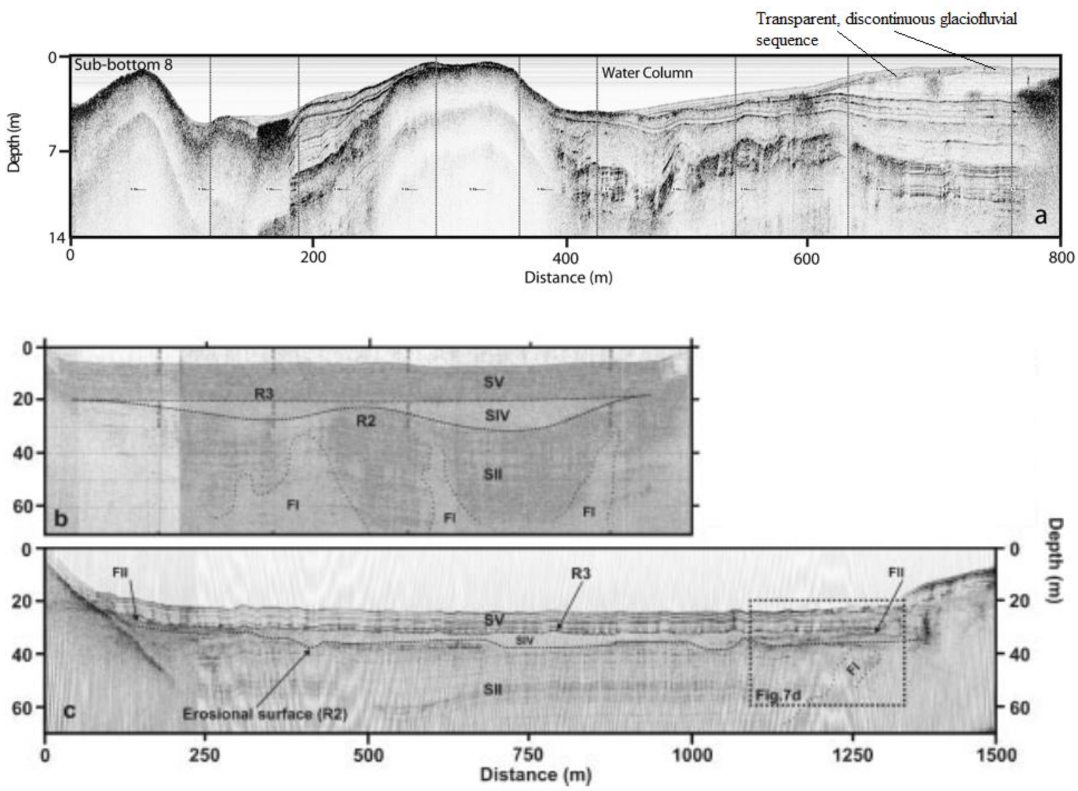


Figure 6.4. Comparison of the glaciofluvial sequence within this study (a) to a similar acoustically transparent facies (SIV) interpreted as glaciofluvial outwash in the subsurface of Loch Ness, Scotland (Turner *et al*., 2012). The position of the glaciofluvial sequence within this study is also similar to

that observed within Loch Ness, with glaciofluvial sedimentation occurring above glaciolacustrine deposition and below a Holocene drape facies.

Figure 6.5. Example of glaciofluvial sedimentation occurring above glaciolacustrine deposits during ice retreat for an exposure adjacent to Lake Pukaki, New Zealand (Glaciers Online, 2008).

6.2.7. Facies LF6: Holocene drape (Lake Infill)

LF6 is interpreted as the uppermost Holocene lake infill facies. This is interpreted because of very small, centimetre-scale reflections which potentially illustrate autochthonous sediment accumulation owing to deposition within the water column. The thin nature of this facies (< 1 m) also illustrates a gradual accumulation of fine fraction material. This facies occurs above all other facies, and drapes all lake-floor geomorphology. The parallel internal reflections which occur continuously through this facies represent changes in sediment properties due to variations in sediment and water flux into the lake. Thinning of this unit downslope may reflect postglacial reworking of unstable, unconsolidated sediments. The sharp transition between this lithofacies and underlying facies is interpreted to have occurred as a result of change in deposition from glacial origin to postglacial infill.

The interpretation of this lithofacies type is supported by similar interpretations from Lake Windermere (Pinson *et al*., 2013) and Loch Ness (Turner *et al*., 2012). In both of these cases, a relatively shallow (2 – 4 m) lithofacies of medium to high-amplitude reflectors occurs that are positioned on the sediment-water boundary (Figure 6.6). Similar facies have also been observed during numerous geophysical investigations of lake sediments (Waldmann *et al*., 2010; Fiore *et al*., 2011).

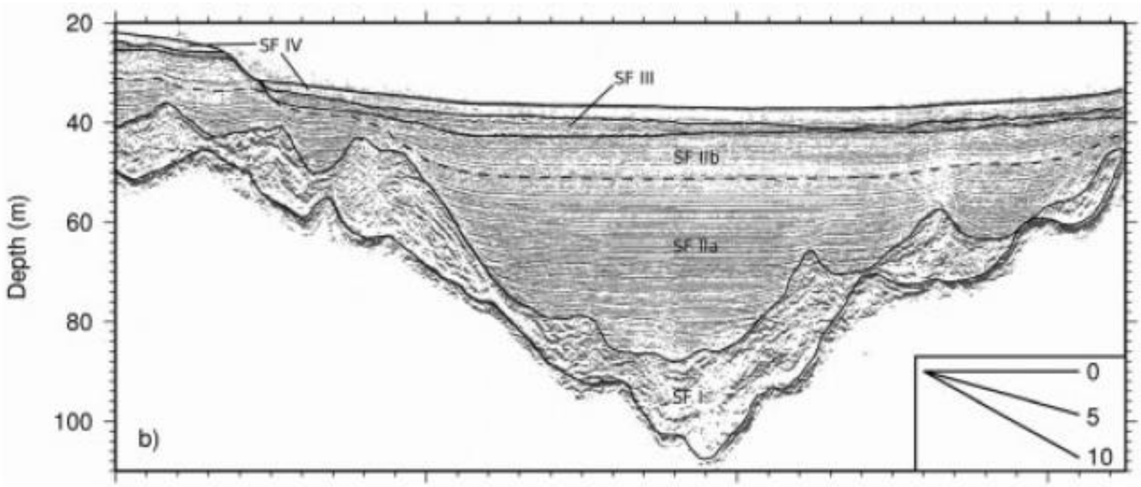


Figure 6.6. A Holocene, ~ 0 – 11 ka BP lithofacies interpreted by Pinson *et al*. (2013) for Lake Windermere, Lake District (SF IV facies).

6.2.8. Loss of detail

In all cases, the Holocene drape facies occurs above this lack of detail. A loss of detail is apparent owing to random, sharp changes from detailed imagery to an initially strong reflected facies which becomes increasingly transparent with depth. Such sharp lateral changes have also been recorded at other locations in the Lake District. Pinson *et al*. (2014) noted such loss of detail, with similar internal facies characteristics to those reported here. This resulted in large portions of Lake Windermere being inaccessible for sub-surface analysis. Pinson *et al*. (2014) suggested such a loss of detail to result from locations where shallow gas was present in the stratigraphy.

6.3. Sub-bottom profile interpretations

6.3.1. Sub-bottom Profile 1

Sub-bottom 1 (Figure 5.2; 5.6) shows an undulating bedrock or till facies, potentially sculpted by previous glacial erosion, which is subsequently infilled with retreat phase, glaciolacustrine sediment. The bedrock or till facies within this transect forms smooth undulations on the lake floor. This is then overlain by high reflected lower glaciolacustrine sediments. The less frequent internal stratification and the high reflection of LF1 may potentially illustrate the compaction or deformation of the internal properties LF2. This could indicate the presence of an overriding ice margin or interaction between the ice margin and this facies during retreat. Following this, lake-infill deposits drape the entire lake-floor. This sub-bottom transect also illustrates potential dropstone features (see Figure 6.10), which occur at ~ 100 m across the transect. As at other locations, LF2 drapes the underlying facies with LF3 then infilling depressions within the bedrock or till. The consistent drape of LF2 supports deposition across a wide swathe of the lake, resulting in an even deposition. The infilling of bedrock / till depressions by LF3 potentially illustrates more distal input.

6.3.2. Sub-bottom Profile 2

Sub-bottom profile 2 (Figure 5.2; 5.7) is limited in extent but does show an undulating LF1 surface which is overlain by two glaciolacustrine sequences (LF2 and LF3). As within Sub-bottom 1, till or bedrock undulations are draped by LF2, which may have been compacted or internally deformed during deposition at the ice margin. Lonne (1995) reported syndepositional, glaciotectonic deformation occurring within glaciolacustrine facies due to ice-front oscillations from a range of fjordal basins in Norway and Svalbard. The infilling of LF2 by LF3 within this transect at ~ 550 m is interpreted to have resulted due to an initial incision into LF2. This could have resulted from an increase in meltwater flux at the ice margin, with a meltwater plume or subglacial drainage event resulting in incision into LF2. This hypothesis further emphasises a proximal deposition for LF2. Once this incision has occurred, a move to distal deposition results in an infilling of this area by LF3 from suspension settling of glaciolacustrine material.

6.3.3. Sub-bottom Profile 3

Sub-bottom 3 (Figure 5.2; 5.8) shows either an undulating bedrock or possibly two moraine, till-filled features, subsequently draped in a lower, partially stratified glaciolacustrine sequence (LF2). This is then overlain by LF3. The differences in internal reflectance and lithofacies thickness for LF2 and LF3 facies is illustrated well within this transect. LF2 has a high reflectance whereas LF3 shows a large variation between transparent and strong internal reflections. This is interpreted to result from changes in grain size, with the upper glaciolacustrine facies illustrating large variations in grain size in comparison to the lower facies. This may also illustrate a potential for compaction within the lower glaciolacustrine unit resulting from ice marginal interactions and glaciotectonic deformation (Lonne, 1995). This is further emphasised by the presence of clear stratification within LF2 within some locations (Sub-bottom 5) and very little stratification in other locations (for example, this transect).

6.3.4. Sub-bottom Profile 4

Sub-bottom 4 (Figure 5.2; 5.9) contains a submerged moraine on the floor of Bassenthwaite Lake and therefore provides evidence for LF1 being potentially composed of till. The difference in water depth either side of this feature potentially illustrates sedimentation occurring at the glacier snout, beyond the moraine with subsequent recession interpreted to occur rapidly owing to a lack of outwash material. The potential moraine feature is composed of LF1, with LF2 draping this high angle, steep undulation. Previous interpretations within this study of deformation occurring within LF2 may illustrate an overriding ice mass deforming LF2 over this bed undulation and incorporating this facies within the moraine. Following this, LF3 onlaps and infills the areas adjacent to this.

Mager and Fitzsimons (2007) identified moraines within the Tasman Valley, New Zealand that were also composed of glaciolacustrine sediments. Within the Tasman Valley a moraine formed during a prior glaciation resulted in further sedimentation occurring around this (Figure 6.7). A retreating ice mass would result in glaciolacustrine sedimentation around the moraine with the sedimentary architecture proposed by Mager and Fitzsimons similar to that observed within Sub-bottom 4.

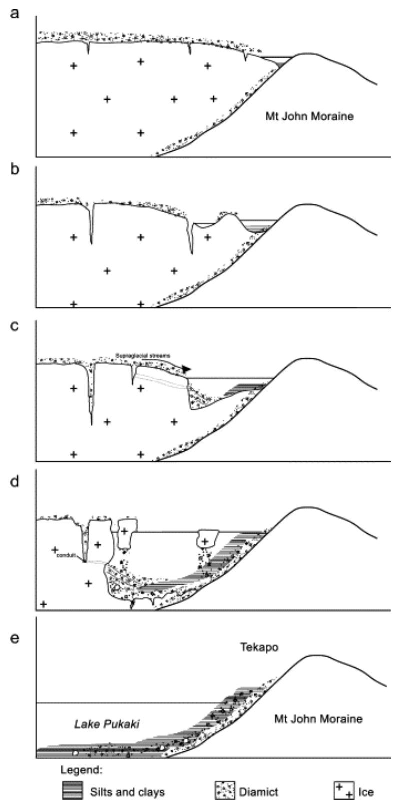


Figure 6.7. Glaciolacustrine sedimentation occurring adjacent to and over a moraine and illustrating similar architectural characteristics to the moraine observed within Bassenthwaite Lake (Mager and Fitzsimons, 2007).

6.3.5. Sub-bottom Profile 5

Sub-bottom 5 (Figure 5.2; 5.10) illustrates a relatively deep buried bedrock or till surface with thick sequences of overlying LF2 and LF3 deposits. A sharp transition from LF2 to LF3 is interpreted to represent an abrupt change in depositional environment. This has been outlined in earlier sections and potentially illustrates a quick transition from ice-proximal to ice-distal deposition. A transition is also apparent within LF3 from transparent facies to increasingly more strong reflections with increasing proximity to the lake floor surface. This is interpreted to reflect changes in sediment size with a gradually retreating ice mass, resulting in finer particle size deposition. LF3 also appears eroded at the lake surface, owing to the non-continuous nature of the internal stratification. This is interpreted here to have resulted from deglacial erosion, potentially related to meltwater channel incision.

Such changes in glaciolacustrine reflectance resulting from changes in sediment flux have also been identified by Heirman *et al*. 2011 within the sub-surface sediments in Lago Puyehue, Chilean Lake District. Thinner beds within the glaciolacustrine sequences were interpreted to illustrate slower, less pulsed sedimentation with a finning upwards resulting from waning sediment supply and increasing distance from the sediment source (Heirman *et al*., 2011).

6.3.6. Sub-bottom Profile 6

Sub-bottom 6 (Figure 5.2; 5.11) is similar to Sub-bottom 5 (see Section 6.3.5). This profile includes a deeply buried scoured bedrock that is subsequently overlain by glaciolacustrine deposits, varying in thickness and reflectance properties. The change in reflectance within LF3 is interpreted again as changes in grain size relating to ice front retreat (Heirman *et al*., 2011; Pinson *et al*., 2013).

6.3.7. Sub-bottom Profile 7

Sub-bottom 7 (Figure 5.2; 5.12) contains a potential drumlin feature which is preconditioned by the underlying bedrock or till facies. The bedrock or till facies has a rounded and smoothed upper bounding surface. The undulating surface is then draped by a lower glaciolacustrine facies, which does not change in thickness and illustrates high reflectance. The high reflectance of this facies and the relatively sparse stratification within LF2 in this transects potentially illustrates compaction and plastering of the unit across the undulating till or bedrock surface resulting from an overriding ice mass. Following this lower glaciolacustrine sedimentation, a retreating ice margin results in the infilling of depressions within Sub-bottom 7 and a subduing of the topography in the inter-drumlin areas.

Therefore the drumlin feature has been preconditioned by bedrock or till topography with subsequent sedimentation draping and infilling the bedform. Glaciolacustrine sediments have been observed within other drumlins, however primarily occur between till units (Kerr and Eyles, 2007).

6.3.8. Sub-bottom Profile 8

Sub-bottom 8 (Figure 5.2; 5.13) is the longest detailed subsurface transect obtained. This transect contains all lithofacies observed beneath Bassenthwaite Lake. LF1 illustrates a smoothed surface and is interpreted to illustrate an eroded, scoured surface. This has also been observed for other lower till and bedrock units at other locations and has been attributed to prior glacial erosion (Waldmann *et al*., 2010; Pinson *et al*., 2013). As in Sub-bottom 7, the lower glaciolacustrine unit overlies LF1 and is partially internally stratified in some locations and illustrates a high reflectance. The high reflectance may illustrate a compaction and plastering of LF1 across the underlying till or bedrock surface resulting from an overriding ice mass. LF2 within this transect does not show the consistent internal stratification as it does in other locations (Sub-bottom 5). This could potentially illustrates that LF1 may be deformed within this transect, resulting in the absence of continuous internal stratification. Sub-bottom 8 illustrates the preferential infilling of lake-floor depressions by LF3 and is a similar depositional environment to that observed by Waldmann *et al*. (2010) in Lago Fagnano, southern Patagonia. Waldmann *et al*. (2010) report thicker accumulations of glaciolacustrine sediment within depressions in the acoustic basement and thinner deposits on lake-floor undulations (Figure 6.8). Waldmann *et al*. (2010) also report a moraine composed of basement till (similar to the moraine within Sub-bottom 4) and glaciolacustrine sediments. A glaciofluvial sequence, LF5, is observed above both LF2 and LF3. A similar lithofacies was described by Turner *et al*. (2012) with glaciofluvial sediments overlying glaciolacustrine sediments and linked to abrupt subglacial meltwater evacuation at an ice margin. This abrupt release of meltwater could have resulted in the erosional incision of LF2 observed in Sub-bottom 5.

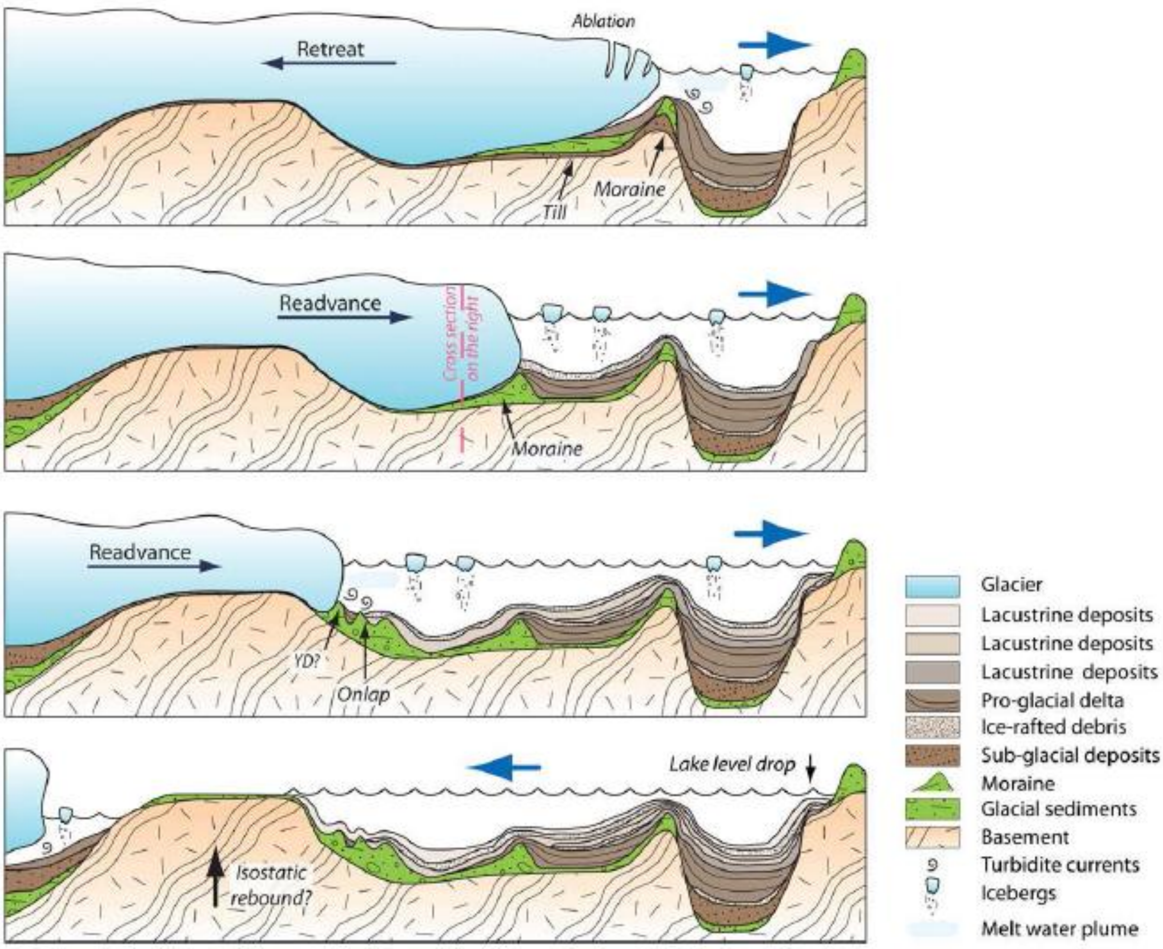


Figure 6.8. Depositional model for Lago Fagnano, southern Patagonia, illustrating basement undulations, moraine features and glaciolacustrine infilling of basement depressions. These are also observed within the sub-surface of Bassenthwaite Lake.

6.3.9. Drumlin facies interpretation

The orientation and proximity of the sub-bottom profiles allow a lateral interpretation of sub-surface lithofacies, structures and details. The feature identified in Sub-bottom 7 and Sub-bottom 8 (Figure 5.12; 5.13; 5.14) is interpreted as a drumlin (Figure 6.9). Firstly, it is adjacent to a non-submerged drumlin adjacent to the Lake. Secondly, this feature is ~ 300 m in length, ~ 50 m in width and ~ 7 m in height and conforms to average drumlin sizes and elongation ratios as measured in other studies (Clark *et al*., 2009). The drumlin also conforms to the criteria of a smooth, streamlined feature that resembles 'an egg half-buried along its long axis' (Clark *et al*., 2009). The core of the drumlin is composed of LF1 which is interpreted as either bedrock or till. Such bedrock and till 'cored' drumlins are common within the literature (Stokes *et al*., 2011). Furthermore, other drumlins have been identified with glaciolacustrine sediments within them (Kerr and Eyles, 2007).

The undulating LF1 bedrock or till surface, which has been previously glacially eroded, occurs within both transects within the 2 adjacent images. LF2 occurs, in both transects, as a high reflectance glaciolacustrine sequence, that illustrates weak and partial stratification and has been interpreted to result from compaction and plastering of the lithofacies, over the underlying undulating surface. This may have resulted from a readvance of ice and therefore 'drumlinisation' of the glaciolacustrine sediments over the underlying basement. LF3 subsequently infills the depressions within the basement till or bedrock and changes in reflectance upwards within the lithofacies. LF3 occurs thinly over the drumlin feature and this upper glaciolacustrine sediment infills the inter-drumlin area, resulting in a more subdued topography. Following glaciolacustrine deposition a relatively thin re-advancing till unit is interpreted to have occurred and also becomes incorporated into the drumlin feature.

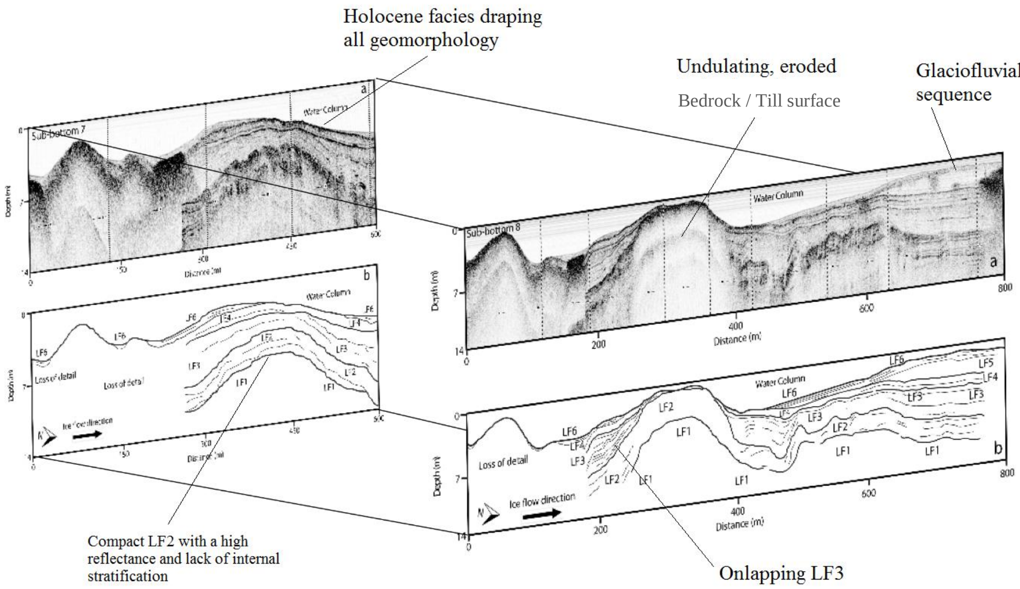


Figure 6.9. Interpretation of a drumlin feature, illustrating a lateral continuation of sedimentary architecture and an undulating bedrock or till surface resulting in the lake-floor drumlin geomorphology.

6.3.10. Interpretation of small-scale structures and details

Small-scale structures and details observed in sub-bottom profiles are interpreted here. A linear ridge of centimetre-scale reflections was observed in Sub-bottom 1 (Figure 5.6). These reflections occur within different stratigraphic layers and are interpreted here as dropstones (Figure 6.10). Dropstones are common within glaciolacustrine deposits (Thomas *et al*., 1985) and have also been observed to the north of Bassenthwaite Lake within a proglacial lake in the Solway Lowlands by Livingstone *et* *al*. (2010). Relatively small, ~ 50 cm high features are observed to protrude into the Holocene drape in Sub-bottom 3 at ~ 600 m (Figure 6.11). These are interpreted as boulders deposited on the lake floor. As the ice margin would have long retreated when these features were deposited, these boulders may reflect either an input of adjacent valley wall material or the re-deposition of glacial material that was deposited above the lake surface. Large variations in reflectance within LF3 within Sub-bottom 7 illustrates an alteration in sediment size. This may result from changes in seasonal meltwater and sediment input and an increasing distance between the ice mass and lake deposition. A small lens-shaped feature in LF3 within Sub-bottom 5 is interpreted as an ice-berg dump or grounding structure (Figure 6.12). The conical mound is typical. This interpretation follows similar observations of ice-berg grounding features within Scotland (Thomas *et al*., 1985) and in fjordal basins in Norway and Svalbard (Lonne, 1995). Lonne (1995) observed a 15 m long block of diamicton, of a lens shape, deposited in glaciolacustrine sediments (Figure 6.12) which also illustrated remobilisation of sediments once deposited by the iceberg.

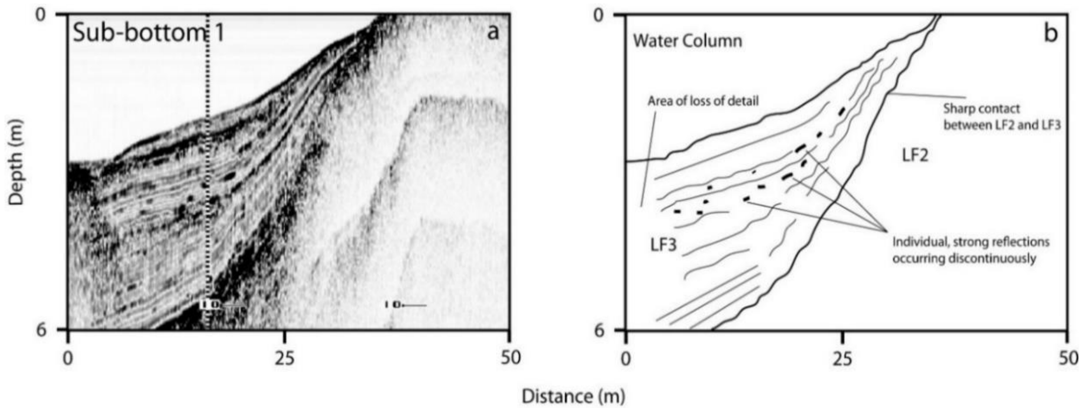


Figure 6.10. Example of sub-metre, discontinuous strong reflections within LF3 in Sub-bottom 1 and interpreted as dropstones.

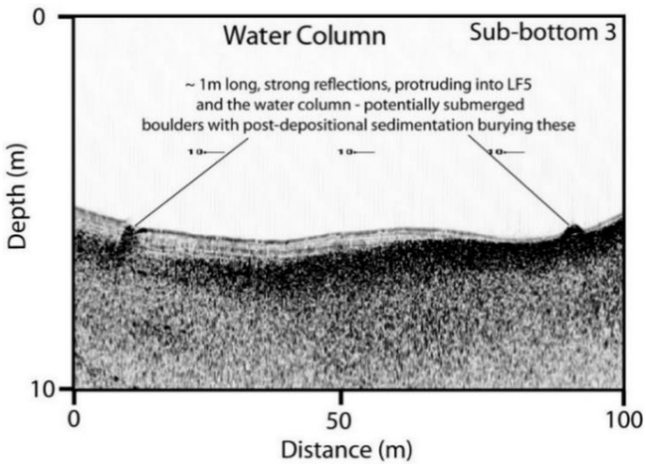


Figure 6.11. Two lake-floor features illustrating strong reflections and protruding into LF6 deposits and interpreted as erratic boulder material.

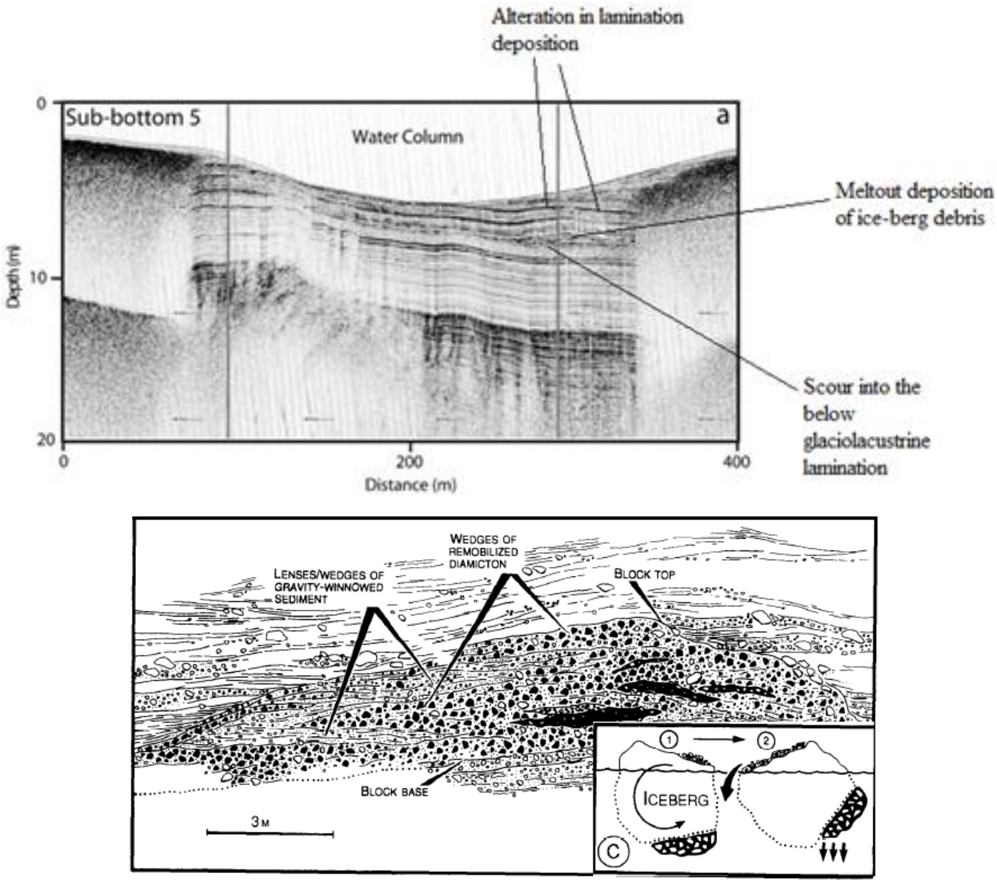


Figure 6.12. Ice-berg grounding area with meltout debris resulting in subsequent glaciolacustrine laminations draping this feature (above). Example of an ice-rafted debris lens in an ice-contact submarine fan, illustrated by Lonne (1995) (below). An inset (c) of the common style of sediment deposition from a drifting iceberg is also provided.

6.4. Summary of interpretations

The analysis of geomorphology adjacent to Bassenthwaite Lake in addition to both geomorphology and sedimentology obtained from sub-bottom analysis allowed the investigation of basin wide characteristics. An undulating bedrock / till surface underlies all of Bassenthwaite Lake and has been potentially eroded and smoothed by previous glacial abrasion. Following this, the presence of a retreating glacier resulted in the deposition of proximal glaciolacustrine sediments which drape the underlying, undulating bedrock / till. As ice retreat continued, sediment input became more distal, resulting in a thicker sequence of upper glaciolacustrine sediments. However, an input of ice-berg debris is apparent and is illustrated through the presence of ice-berg grounding structures. The presence of glaciofluvial material points to meltwater evacuating the ice margin and depositing local depot centres of well sorted fluvial sediment. All glacial lithofacies are then draped by a Holocene lake-infill.

6.5. The glacial geomorphology of Bassenthwaite Lake and adjacent areas

During the Last Glacial Maximum (LGM) the Bassenthwaite valley would have experienced complete ice coverage with ice emanating from a Lake District ice cap feeding valley outlet glaciers (Evans *et al*., 2005; Hubbard *et al*., 2009). These valley glaciers then fed a larger ice stream which existed in the Solway Lowlands and flowed into the northern Irish Sea Basin (Hubbard *et al*., 2009). This ice stream represents one of the largest outlets of the British-Irish Ice Sheet. Interpretations of ice flow through the Bassenthwaite basin and into a lowland ice stream are supported by drumlins feeding the southern end of the basin. These are observed within this study and also by Livingstone *et al*. (2008) and Hughes *et al*. (2010). As the ice emerges into the lowland area to the north of Bassenthwaite Lake, the long axis of subglacial bedforms migrate towards the west, illustrating a transition to ice flow into the north Irish Sea Basin (Livingstone *et al*., 2008; Hughes *et al*., 2010). This period of maximum ice coverage is also interpreted to have potentially resulted in the erosion of the lowest lithofacies (LF1) surface, producing an eroded, undulating bounding surface.

Ice retreat through the Bassenthwaite basin follows a transition from LGM ice domes to independent ice caps that feed valley glaciers (Hughes *et al*., 2010; Clark *et al*. 2012). The presence of multiple moraine ridges along the Bassenthwaite Lake basin illustrates a stepped retreated throughout the valley. The presence of deformation within some sedimentary facies illustrates a dynamic retreat, with the potential for relatively large, ice-marginal fluctuations. This is evident through the interpretation small-scale deformation occurring within ice-marginal glaciolacustrine sediments, which has also been documented within other ice-marginal, glaciolacustrine units (Lonne, 1995). The presence of glaciolacustrine sediments and ice-berg debris illustrate the potential for changed processes operating within the Bassenthwaite basin. The development of a proglacial lake may result in internal, catchment properties influencing glacier retreat with calving processes and proglacial lake development likely to modulate the glacier from external climate forcing (Warren and Aniya, 1999). Changes in calving dynamics of a valley glacier have been show to reverse the trend expected from climatic controls (Warren and Aniya, 1999). The transition to finer, more thinly bedded glaciolacustrine sediments observed within sub-surface transects illustrates an increasingly more distal ice mass and deglaciation of the Bassenthwaite basin. An input of glaciofluvial material, however, overlying all glaciolacustrine sequences, illustrates an abrupt input of subglacial meltwater during final retreat. This was also observed by Turner *et al*. (2012) within the sub-surface of Loch Ness, Scotland. The presence of drumlins to the north of Derwentwater (located to the south of Bassenthwaite Lake) and also within the lake, protruding as islands, illustrates ice retreat southwards, through this area and into the central Lake District. It is clear therefore that the Bassenthwaite basin operated as a major channel for an ice cap fed valley glacier which eventually joined a lowland ice stream.

Following entire deglaciation of the Bassenthwaite basin, a return to colder conditions during the Younger Dryas resulted in glaciation of some areas of the Lake District (Brown *et al*., 2013) and also resulted in alterations in sedimentation and sub-aerial erosion. Glacier formation and advance during the Younger Dryas is limited to the central Lake District and the Skiddaw range (to the east of Bassenthwaite Lake). A reconstruction of such Younger Dryas extent (Brown *et al*., 2013) illustrates glaciers moving into the southern basin of Derwentwater. Once deglaciated, the modification of the Bassenthwaite basin occurs through fluvial incision of the valley sides and floor and subaerial erosion. Following deglaciation Bassenthwaite Lake is also infilled with an internal lake sediment which was identified within the sub-surface transects draping the entire lake-floor.

6.6. The glacial sedimentary evolution of Bassenthwaite Lake: depositional model

The creation of a depositional model for the sub-surface sedimentary facies within Bassenthwaite Lake (Figure 7.1) extends the analysis of Lake District lake-floor sediments, following investigations of Lake Windermere by Pinson *et al*. (2013). This study provides the first model of ice retreat through the Bassenthwaite basin. The following key stages of lake-floor evolution are proposed:

Stage 1: The lake-floor of the Bassenthwaite basin would have been completely inundated with ice, which extended beyond the basin, feeding a lowland ice stream. This maximum ice coverage provides the conditions for either the deposition of a lowest till unit, which is then eroded at the surface, or the erosion of a pre-existing rock surface (Figure 7.1). This follows numerous other examples of basement erosion which were also observed through the geophysical investigation of glacial retreat basins (Waldmann *et al*., 2010; Heirman *et al*., 2011; Turner *et al*., 2012; Pinson *et al*., 2014).

Stage 2: Ice retreats through the Solway lowlands, to the north of Bassenthwaite Lake and results in a transition from an ice stream feeding glacier to a constrained, retreating valley glacier. The retreat of the glacier within the Bassenthwaite basin results in ice-marginal, glaciolacustrine deposition which is potentially overridden and deformed in localised areas, illustrating small, ice-marginal fluctuations in the retreating glacier (Figure 7.1). This follows other observations of small-scale, ice-marginal fluctuations recorded in glaciolacustrine deposits (Lonne, 1995).

Stage 3: The gradual retreat of the glacier located within Bassenthwaite basin is illustrated by a more distal input of glaciolacustrine material with ice-berg rafted debris identified within sub-bottom transects (Figure 7.1). The transition to a more distal sediment flux is also illustrated through the change in reflectance within glaciolacustrine sediments, with a two-tiered sequence of glaciolacustrine sediments identified. This was also observed by Fiore *et al*. (2011) and Pinson *et al*. (2013) within deglacial basins and was also interpreted to illustrate a change to more distal deposition, resulting from a retreating glacier. The occurrence of a glaciofluvial sequence above such low energy glaciolacustrine facies however indicates the potential for a distal 'pulse' of meltwater, potentially resulting from the abrupt release of stored meltwater within a retreating ice mass.

Stage 4: A re-advancing glacier is identified through the presence of a re-advance till unit which illustrates large-scale glacier advance within localised area of Bassenthwaite Lake. This re-advance unit is also incorporated into the lake-floor drumlin and potentially illustrates a secondary drumlinisation event. The advancing glacier margin provides optimum conditions for coarser deposits over the previous fine-grained glaciolacustrine sediments.

Stage 5: Final retreat of the Bassenthwaite Lake ice mass is illustrated through an input of coarse material (glaciofluvial deposit). This illustrates abundant meltwater resulting from glacier melt and a retreating glacier.

Stage 6: Following entire deglaciation and a cessation of glaciolacustrine and lacustrine sedimentation, a postglacial infill of Holocene, autochthonous material drapes the entire lake floor (Figure 7.1). This facies subdues the lake-floor geomorphology. This stage in lake-floor sedimentary evolution can be correlated to changes in lake-adjacent geomorphology, with a transition to subaerial fluvial erosion occurring.

The depositional model for the sub-surface facies within Bassenthwaite Lake illustrate similarities to those observed elsewhere. These include a geophysical investigation of Lago Fagnano, southern Patagonia by Waldmann *et al*. (2010). The depositional model for Lago Fagnano includes a pre-eroded bedrock from prior glaciations, which later results in the preferential accumulation of glaciolacustrine sequences. Waldmann *et al*. (2010) also observed moraines which were subsequently overlain and draped by glaciolacustrine facies. A depositional model provided by both Pinson *et al*. (2013) and Fiore *et al*. (2011) also illustrates two glaciolacustrine sequences which were attributed to a retreating ice mass and therefore a change in sediment flux. Such alterations in glaciolacustrine flux are also interpreted within Bassenthwaite Lake and illustrate a gradual retreat with no large-scale glacier advances.

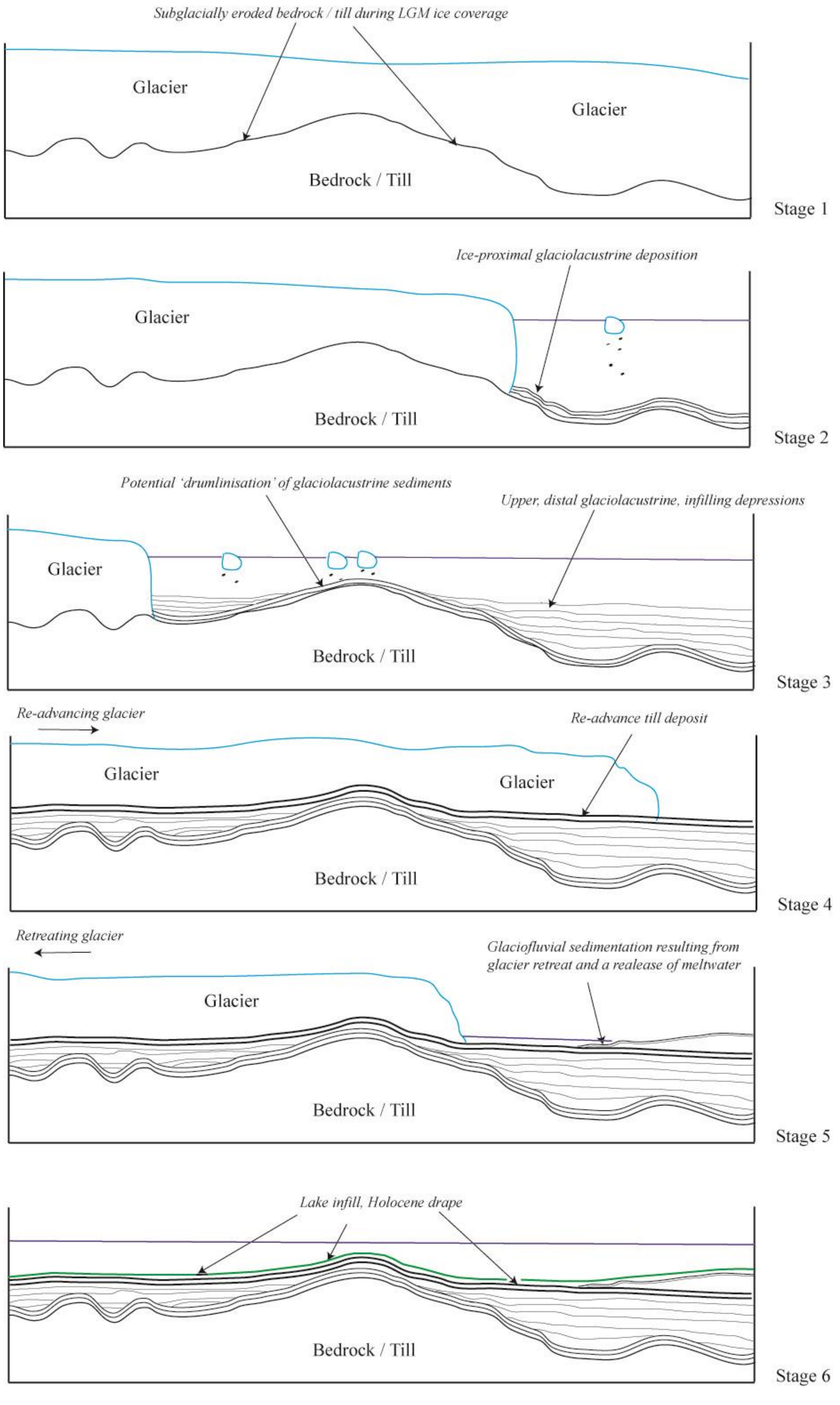


Figure 7.1. Depositional model for the sub-surface facies observed within Bassenthwaite Lake.

6.7. Implications for drumlin formation

The drumlin observed on the floor of Bassenthwaite Lake contains either a core of bedrock or till and is subsequently overlain by a deformed lower glaciolacustrine sequence. This feature is then infilled by distal glaciolacustrine deposits which form thick accumulations on the slopes of the drumlin. The drumlin is finally capped with a thin veneer of till, resulting from a secondary advance of the Bassenthwaite glacier. Therefore, all internal facies occur as conformable sequences within the drumlin. Small-scale oscillations in the ice margin are interpreted to have resulted in the deformation of the lower glaciolacustrine unit over the underlying bedrock or till. The eroded, undulating surface of the lowest till or bedrock unit is of interest as, if interpreted as a till facies, it could potentially illustrate an unstable till bed. These undulations may have formed during total ice coverage with the coupled flow of ice and sediment having the potential to spontaneously create bedforms (Stokes *et al*., 2013). This hypothesis for the lower basement unit forming spontaneous landforms during coupled ice and sediment flow is supported by the presence of multiple undulating features, varying in amplitude and length and located adjacent to each other. It is interesting that other investigations have observed such undulating basement facies (Waldmann *et al*., 2010; Heirman *et al*., 2011; Pinson *et al*., 2013), which have been interpreted as till, but have in all cases been interpreted to have resulted from glacial erosion as opposed to bed perturbations resulting from coupled ice-sediment flow. Regardless of the underlying basement conditions, ice marginal fluctuations of the Bassenthwaite glacier are interpreted to have resulted in the plastering and deforming of the lowest glaciolacustrine unit within the drumlin with partial deformation also recorded in other drumlins (Stokes *et al*., 2013). This lower glaciolacustrine facies which overlies the bedrock or till unit however may represent a ‘veneer’, which is distinct from the drumlin forming process. Veneers have been reported for glaciofluvial and till sediments (Menzies and Brand, 2007; Stokes *et al*., 2013) however glaciolacustrine facies have not been documented. Therefore, the following model is proposed for drumlin formation:

Streamlining of the underlying Skiddaw Slate bedrock or lower basement till during LGM flow phasing produces the undulating drumlin geomorphology which has been subsequently overlain and plastered with a deformed glaciolacustrine veneer. Subsequent distal glaciolacustrine sediments infill depressions within the feature and act to bury the bedrock / till core. Following these glaciolacustrine events, a secondary advance occurs and results in a thin veneer (~ 2 m) of till deposition which caps the drumlin feature.

It is clear, therefore, that the drumlin internal composition and structure results from multiple proglacial sedimentation events and subsequent deformation of these. Similar interpretations of drumlin formation have been proposed by Boyce and Eyles (1991) for the Peterborough drumlin field, central Canada. Thin till veneers were identified by Boyce and Eyles (1991) in addition to the presence of overridden proglacial glaciolacustrine and glaciolacustrine sediments.

6.8. Application of sub-bottom profiling to drumlins

The application of sub-bottom profiling follows the investigation of drumlins using other geophysical methods (Hiemstra *et al*., 2011; Spagnolo *et al*., 2014). Sub-bottom profiling has illustrated its potential for application to submerged drumlins within this study. Repeated profiles can allow an interpretation of adjacent sub-surface composition and structure and provides valuable insights into the sub-surface properties. The relative ease of use also allows multiple profiles to be established, at little cost. The real-time analysis of the sub-surface also allows specific sub-surface locations to be 'targeted' during analysis. This is considered a great advantage of using this technique as brief, in-field interpretations and decisions can be made.

However, it is recognised that the application of sub-bottom profiling did not produce detailed sedimentology at all locations. This is interpreted to have potentially occurred as a result of trapped gas within the upper sediment layers. Following the application of sub-bottom profiling here, there is an apparent need for prior, detailed bathymetry so as to allow the determination of multiple drumlin features. This study therefore provides an insight into the application of sub-bottom profiling, with numerous positive outcomes outlined.

Following the application here, a potentially consideration of other submerged drumlins may be undertaken. These include locations of known submerged drumlins such as Lake Ontario (Kerr and Eyles, 2007), Irish Sea Basin (Van Landeghem *et al*., 2008) and numerous continental shelves, including Antarctica. However, numerous other submerged drumlins and drumlin fields exist. Other subglacial and submerged features however could also be considered with a need to understand mega-scale glacial lineation and other subglacial internal compositions to attempt to understand subglacial processes. Further to application to subglacial features, it is apparent, as from the application of sub-bottom profiling in this study and in other investigations (Waldmann *et al*., 2010; Fiore *et al*., 2011; Turner *et al*., 2012; Pinson *et al*., 2014) that this geophysical technique can provide great insights into Lateglacial retreat dynamics. These include the composition and location of moraines, sedimentary outwash and post-glacial infill.

8. CONCLUSION

The investigation of the sub-surface sedimentology within Bassenthwaite Lake was undertaken in this study and used sub-bottom profiling to assess Lateglacial retreat dynamics. This was combined with geomorphological mapping of the Bassenthwaite Lake basin to understand basin-wide retreat characteristics. During sub-surface analysis, the composition and structure of a drumlin was also obtained and a sedimentary investigation of 2 adjacent transects was undertaken. This investigations represents the first sedimentary and geomorphological analysis of the Bassenthwaite basin and is also the first application of sub-bottom profiling to submerged drumlins. It is clear that the Lateglacial retreat landscape offers important insights into the retreat behaviour of the terrestrial portion of the British-Irish Ice Sheet. The investigation undertaken within the Bassenthwaite basin allows the following key conclusions to be drawn:

1. The Bassenthwaite basin was as a key drainage channel throughout the Pleistocene. Advancing ice was channelled into the southern end of the basin and is interpreted by the convergence of drumlins into the topographically constrained Bassenthwaite basin (Livingstone *et al*., 2008; Hughes *et al*., 2010). The glacier would have then fed a lowland ice stream which drained into the north Irish Sea Basin and is interpreted through the migration of subglacial bedform long-axes towards the west as the ice enters the Solway Lowlands (Livingstone *et al*., 2008; Hughes *et al*., 2010). These ice flow phases throughout the Pleistocene provide the conditions for erosion of the underlying basement or till unit.

2. Following a transition from large-scale ice sheet to localised ice cap dynamics, deglaciation of the Bassenthwaite basin occurred. A punctuated retreat of the glacier through the Bassenthwaite basin is apparent and is interpreted through the presence of numerous moraine ridges identified through geomorphological mapping. Ice-marginal glaciolacustrine sedimentation also occurred into a proglacial lake with small, localised, ice-marginal fluctuations resulting in the deformation of some glaciolacustrine sediments.

3. Continued glacier retreat is illustrated by the accumulation of thick sequences of upper glaciolacustrine sediments, which are distinct from the underlying glaciolacustrine unit due to changes in grain size. Ice-berg grounding areas were also identified with distal glaciolacustrine sediments infilling depressions on the lake-floor. The two-tiered glaciolacustrine sediments, resulting from a transition from ice-proximal to ice-distal deposition has also been recognised within other deglacial lakes (Fiore *et al*., 2011; Pinson *et al*., 2013). An input of glaciofluvial material potentially illustrates a release of stored

meltwater within a retreating ice mass following similar interpretations by Turner *et al*., 2012 within Loch Ness, Scotland. Due to the presence of deformation within the sedimentary some facies a dynamic and punctuated retreat is interpreted, with some large readvances of the ice-margin apparent.

4. Entire deglaciation of the Bassenthwaite basin subsequently occurred and is illustrated by a transition to an input of fine-grained, lake-infill sediments associated with internal lake sedimentation. This facies drapes the entire lake floor. The transition to entire basin deglaciation is also represented by the presence of fluvial incised basin walls and floor and an increase in subaerial erosion illustrated by the presence of scree-slope material.

5. During the sub-bottom investigation of Bassenthwaite Lake the internal composition and structure of a drumlin was identified. This drumlin is composed of a till or bedrock core and is subsequently overlain by a veneer of deformed lower glaciolacustrine sediments. Following this, distal glaciolacustrine sediments infill depressions within the feature. A secondary advance of the glacier results in the accumulation of another sedimentary package, a re-advance till. The drumlin therefore reflects numerous ice-marginal fluctuations and deformational events resulting in drumlinisation of proglacial material.

6. The application of sub-bottom profiling provided the internal composition and structure of a drumlin. Therefore, this technique has the potential to consider the sub-surface characteristics of submerged drumlins at other locations. The identification of numerous submerged features prior to investigation, however, is recommended and the potential for a loss in the sub-surface detail is recognised. Sub-bottom profiling, however, again illustrates its applicability in considering deglacial sediments and follows the application of this technique to other deglaciated lake basins (Fiore *et al*., 2011; Turner *et al*., 2012; Pinson *et al*., 2013).

9. REFERENCES

Alley, R. B., Clark, P. U., Huybrechts, P. and Joughin, I. (2005) 'Ice-sheets and sea-level changes.' *Science,* **310** (5747), 456 – 460.

Ballantyne, C. K., Stone, J. O. and Fifield, L. K. (2009) 'Glaciation and deglaciation of the SW Lake District, England: implications of cosmogenic 36 Cl exposure dating.' *Proceedings of the Geologists' Association,* **120** (2), 139 – 144.

Bamber, J. L. and Aspinall, W. P. (2013) 'An expert judgement assessment of future sea level rise from the ice sheets.' *Nature Climate Change,* **3** (4), 424 – 427.

Baranowski, S. (1979) 'The origin of drumlins as an ice-rock interface problem.' *Journal of Glaciology*, **23** (89), 435 – 436.

Benn, D. and Evans, D. J. (2014) *Glacier and glaciation*. Routledge.

Bennett, M. R. (2003) 'Ice streams as the arteries of an ice sheet: their mechanics, stability and significance.' *Earth-Science Reviews*, **61** (3), 309 – 339.

Bennett, M. R. and Glasser, N. F. (2011) *Glacial geology: ice sheets and landforms*. John Wiley & Sons.

Boulton, G. and Hagdorn, M. (2006) 'Glaciology of the British Isles Ice Sheet during the last glacial cycle: form, flow and lobes.' *Quaternary Science Reviews*, **25**, 3359 – 3390.

Boulton, G. S. (1987) 'A theory of drumlin formation by subglacial sediment deformation.' *Drumlin Symposium*.

Boulton, G. S. (1996) 'Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation.' *Journal of Glaciology*, **42** (140), 43 – 62.

Boyce, J. I. and Eyles, N. (1991) 'Drumlins carved by deforming till streams below the Laurentide ice sheet.' *Geology*, **19**, 787 – 790.

British Geological Survey (1999) *1:50,000 Scale Geological Map – Maps 23 and 29*. British Geological Survey.

British Geological Survey (2012) Windermere: lake environment and glacial history. Available at: http://www.bgs.ac.uk/research/highlights/2012/lakeWindermere.html

Brown, V. H., Evans, D. J. A., Vieli, A. and Evans, I. S. (2013) ‘The Younger Dryas in the English Lake District: reconciling geomorphological evidence with numerical model outputs.’ *Boreas*, **10**, 9483.

Burke, M. J., Woodward, J., Russell, A. J., Fleisher, P. J. and Bailey, P. K. (2008) ‘Controls on the sedimentary architecture of a single event englacial esker: Skeidararjoukll, Iceland.’ *Quaternary Science Reviews*, **27** (19), 1829 – 1847.

Clapperton, C. M. (1989) ‘Asymmetrical drumlins in Patagonia, Chile.’ *Sedimentary Geology*, **62**, 387 – 398.

Clark, C. D. (1997) ‘Reconstructing the evolutionary dynamics of former ice sheets using multi-temporal evidence, remote sensing and GIS.’ *Quaternary Science Reviews*, **16** (9), 1067 – 1092.

Clark, C. D., Evans, D. J., Khatwa, A., Bradwell, T., Jordan, C. J., Marsh, S. H. and Batemen, M. D. (2004) ‘Map and GIS database of glacial landforms and features related to the last British Ice Sheet.’ *Boreas*, **33** (4), 359 – 375.

Clark, C. D., Hughes, A. L., Greenwood, S. L., Jordan, C. and Sejrup, H. P. (2012) ‘Pattern and timing of retreat of the last British-Irish Ice sheet.’ *Quaternary Science Reviews*, **44**, 112 – 146.

Clark, C. D., Hughes, A. L., Greenwood, S. L., Spagnolo, M. and Ng, F. S. (2009) ‘Size and shape characteristics of drumlins, derived from a large sample, and associated scaling laws.’ *Quaternary Science Reviews*, **28** (7), 677 – 692.

Clark, C. D., Tulaczyk, S. M., Stokes, C. R. and Canals, M. (2003) ‘A groove-ploughing theory for the production of mega-scale glacial lineations, and implications for ice-stream mechanics.’ *Journal of Glaciology*, **49** (165), 240 – 256.

Clarke, G. K. (2005) ‘Subglacial processes.’ *Annual Review of Earth and Planetary Science*, **33**, 247 – 276.

Cofaigh, C. O., Dowdeswell, J. A., Allen, C. S., Hiemstra, J. F., Pudsey, C. J., Evans, J. and Evans, D. J. (2005a) ‘Flow dynamics and till genesis associated with a marine-based Antarctic palaeo-ice stream.’ *Quaternary Science Reviews*, **24** (5), 709 – 740.

Dardis, G. F. and Hanvey, P. M. (1994) ‘Sedimentation in a drumlin lee-side subglacial wave cavity, northwest Ireland.’ *Sedimentary Geology*, **91**, 97 – 114.

Darvill, C. M., Stokes, C. R., Bentley, M. J. and Lovell, H. (2014) 'A glacial geomorphological map of the southernmost ice lobes of Patagonia: the Bahia Inutil-San Sebastian, Magellan, Otway, Skyring and Rio Gallegos lobes.' *Journal of Maps,* **10** (3), 500 – 520.

Evans, D. J. A., Livingstone, S. J., Vieli, A. and O'Cofaigh, C. O. (2009) 'The palaeoglaciology of the central sector of the British and Irish Ice Sheet: reconciling glacial geomorphology and preliminary ice sheet modelling.' *Quaternary Science Reviews*, **28**, 739 – 757.

Evans, D. J. A., Phillips, E. R., Hiemstra, J. F. and Auton, C. A. (2006) 'Subglacial till: formation, sedimentary characteristics and classification.' *Earth-Science Reviews*, **78** (1), 115 – 176.

Eyles, N., Boyce, J. I. and Barendregt, R. W. (1999) 'Hummocky moraine: sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds.' *Sedimentary Geology,* **123** (3), 163 – 174.

Fiore, J., Girardclos, S., Pugin, A., Gorin, G. and Wildi, W. (2011) 'Wurmian deglaciation of western Lake Geneva (Switzerland) based on seismic stratigraphy.' *Quaternary Science Reviews,* **30** (3), 377 – 393.

Fisher, T. G. and Spooner, I. (1994) 'Subglacial meltwater origin and sub aerial meltwater modifications of drumlins near Morley, Alberta, Canada.' *Sedimentary Geology,* **91**, 285 – 298.

Fowler, A. C. (2000) 'An instability mechanism for drumlin formation.' *Geological Society, London, Special Publications,* **176**, 307 – 319.

Fuller, S. and Murray, T. (2002) 'Sedimentological investigations in the fore field of an Icelandic surge-type glacier: implications for the surge mechanism.' *Quaternary Science Reviews,* **21**, 1503 – 1520.

Glaciersonline (2008) 'Complex interbedding of sand and gravel of glaciofluvial origin, and glaciolacustrine silts.' Available at: http://www.swisseduc.ch/glaciers/new_zealand/tasman_braided_rivers/index-en.html?id=9

Goldstein, B. (1994) 'Drumlins of the Puget Lowland, Washington State, USA.' *Sedimentary Geology,* **91**, 299 – 312.

Graham, A. G. C., Larter, R. D., Gohl, K., Hillenbrand, C. D., Smith, J. A. and Kuhn, G. (2009) 'Bedform signature of a West Antarctic palaeo-ice stream reveals a multi-temporal record of flow and substrate control.' *Quaternary Science Reviews*, **28** (25), 2774 – 2793.

Gray, L., Joughin, I., Tulaczyk, S., Spikes, V. B., Bindschandler, R. and Jezek, K. (2005) 'Evidence for subglacial water transport in the West Antarctic Ice Sheet through three-dimensional satellite radar interferometry.' *Geophysical Research Letters*, **32** (3).

Hart, J. K. (1995) 'Recent drumlins, flutes and lineations at Vestari-Hagafellsjokull, Iceland.' *Journal of Glaciology*, **41**, 596 – 606.

Hart, J. K. (1999) 'Identifying fast ice flow from landform assemblages in the geological record: a discussion.' *Annals of Glaciology*, **28** (1), 59 – 66.

Heirman, K., De Batist, M., Charlet, F., Moernaut, J., Emmanuel, C., Brummer, R., Pino, M. and Urrutia, R. (2011) 'Detailed seismic stratigraphy of Lago Puyehue: implications for the mode and timing of glacier retreat in the Chilean Lake District.' *Journal of Quaternary Science*, **26** (7), 665 – 674.

Heroy, D. C. and Anderson, J. B. (2005) 'Ice-sheet extent of the Antarctic Peninsula region during the Last Glacial Maximum (LGM) – insights from glacial geomorphology.' *Geological Society of America Bulletin*, **117**, 1497 – 1512.

Hiemstra, J. F., Kulessa, B., King, E. C. and Ntarlagiannis, D. (2011) 'The use of integrated sedimentological and geophysical methods in drumlin research-a case study of Pigeon Point, Clew Bay, Northwest Ireland.' *Earth Surface Processes and Landforms*, **36** (14), 1860 – 1871.

Hill, A. R. (1971) 'The internal composition and structure of drumlins in North Down and South Antrim, Northern Ireland.' *Geografiska Annaler*, **53**, 14 – 31.

Hindmarsh, R. C. (1998) 'The stability of a viscous till sheet coupled with ice flow, considered at wavelengths less than the ice thickness.' *Journal of Glaciology*, **44**, 285 – 292.

Hogan, K. A., Dix, J. K., Lloyd, J. M.., Long, A. J. and Cotterill, C. J. (2011) 'Seismic stratigraphy records the deglacial history of Jakobshaven Isbrae, West Greenland.' *Journal of Quaternary Science*, **26** (7), 757 – 766.

Hubbard, A., Bradwell, T., Golledge, N., Hall, A., Patton, H., Sugden, D., Cooper, R. and Stocker, M. (2009) 'Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British-Irish ice sheet.' *Quaternary Science Reviews*, **28**, 758 – 776.

Hughes, A. L. C., Clark, C. D. and Jordan, C. J. (2010) 'Subglacial bedforms of the last British Ice Sheet.' *Journal of Maps*, **6**, 543 – 563.

Johnson, M. D., Schomacker, A., Benediktsson, I. O., Geiger, A. J., Ferguson, A. and Ingolfsson, O. (2010) 'Active drumlin field revealed at the margin of Mulajokull, Iceland; a surge-type glacier.' *Geology*, **38** (10), 943 – 946.

Jorgensen, F. and Piotrowski, J. A. (2003) 'Signature of the Baltic Ice Stream on Funen Island, Denmark during the Weichselian glaciation.' *Boreas*, **32** (1), 242 – 255.

Kerr, M. and Eyles, N. (2007) 'Origin of drumlins on the floor of Lake Ontario and in upper New York State.' *Sedimentary Geology*, **193**, 7 – 20.

King, E. C., Woodward, J. and Smith, A. M. (2007) 'Seismic and radar observations of subglacial bedforms beneath the onset zone of Rutford Ice Stream, Antarctica.' *Journal of Glaciology*, **53**, 665 – 672.

Kjaer, K. H., Kruger, J. and van der Meer, J. J. M. (2003) 'What causes till thickness to change over distance? Answers from Myrdalsjokull, Iceland.' *Quaternary Science Reviews*, **22**, 15 – 17.

Kleman, J. and Hatterstrand, C. (1999) 'Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum.' *Nature*, **402** (6757), 63 – 66.

Knight, J. (2014) 'Subglacial Hydrology and Drumlin Sediments in Connemara, Western Ireland.' *Geografiska Annaler: Series A, Physical Geography*.

Livingstone, S. J., Cofaigh, C. O. and Evans, D. J. (2010) 'A major ice drainage pathway of the last British-Irish ice sheet: the Tyne Gap, northern England.' *Journal of Quaternary Science*, **25** (3), 354 – 370.

Livingstone, S. J., O'Cofaigh, C. and Evans, D. J. A. (2008) 'Glacial geomorphology of the central sector of the last British-Irish Ice Sheet.' *Journal of Maps*, **4**, 358 – 377.

Lonne, I. (1995) 'Sedimentary facies and depositional architecture of ice-contact glaciomarine systems' *Sedimentary Geology*, **98** (1), 13 – 43.

Mager, S. and Fitzsimons, S. (2007) 'Formation of glaciolacustrine Late Pleistocene end moraines in the Tasman Valley, new Zealand.' *Quaternary Science Reviews*, **26** (5), 743 – 758.

McCabe, A. M. and Dardis, G. F. (1989) 'A geological view of drumlins in Ireland.' *Quaternary Science Reviews*, **8** (2), 169 – 177.

Meehan, R. T., Warren, W. P. and Gallagher, C. J. D. (1997) 'The sedimentology of a Late Pleistocene drumlin near Kingscourt, Ireland.' *Sedimentary Geology*, **111** (91), 105.

Menzies, J. (1979) ‘A review of the literature on the formation and location of drumlins.’ *Earth-Science Reviews*, **14** (4), 315 – 359.

Menzies, J. and Brand, U. (2007) ‘The internal sediment architecture of a drumlin, Port Byron, New York State, USA.’ *Quaternary Science Reviews*, **26**, 322 – 355.

Mitchell, W. A. and Riley, J. M. (2006) ‘Drumlin map of the Western Pennines and southern Vale of Eden, northern England, UK. *Journal of Maps*, **2** (1), 10 – 16.

Nenonen, J. (1994) ‘The Kaituri stratigraphy in the Kangasniemi area, Finland.’ *Sedimentary Geology*, **91**, 365 – 372.

O’Cofaigh, C. O. and Stokes, C. R. (2008) ‘Reconstructing ice-sheet dynamics from subglacial sediments and landforms: introduction and overview.’ *Earth Surface Processes and Landforms*, **33** (4), 495 – 502.

O’Cofaigh, C., Dowdeswell, J. A., Allen, C. S., Hiemstra, J. F., Pudsey, C. J., Evans, J. and Evans, D. J. A. (2005) ‘Flow dynamics and till genesis associated with a marine-based Antarctic palaeo-ice stream.’ *Quaternary Science Reviews*, **24**, 709 – 740.

O’Cofaigh, C., Pudsey, C. J., Dowdeswell, J. A. and Morris, P. (2002) ‘Evolution of subglacial bedforms along a palaeo-ice stream, Antarctic Peninsula continental shelf.’ *Geophysical Research Letters*, **29** (8), 1199.

Paterson, W. S. B. (2001) *The Physics of Glaciers*. Butterworth-Heinemann.

Patterson, C. J. and Hooke, L. H. (1995) ‘Physical environment of drumlin formation.’ *Journal of Glaciology*, **41**, 30 – 38.

Pinson, L. J., Vardy, M. E., Dix, J. K., Henstock, T. J., Bull, J. M. and Maclachlan, S. E. (2013) ‘Deglacial history of glacial lake Windermere, UK: implications for the central British and Irish Ice Sheet.’ *Journal of Quaternary Science*, **28** (1), 83 – 94.

Pritchard, H. D., Arthern, R. J., Vaughan, D. G. and Edwards, L. A. (2009) ‘Extensive dynamic thinning on the margins of the Greenland and Antarctic ice sheets.’ *Nature*, **461** (7266), 971 – 975.

Rahmstorf, S. (2002) ‘Ocean circulation and climate during the past 120,000 years.’ *Nature*, **419** (6903), 207 – 214.

Ramsbottom, A. E. (1976) Depth charts of the Cumbrian lakes. Hyperion

Rattas, M. and Piotrowski, J. A. (2003) 'Influence of bedrock permeability and till grain size on the formation of the saadjarve drumlin field, Estonia, under an east-Baltic Weichselian ice stream.' *Boreas*, **32** (1), 167 – 177.

Raunholm, S., Sejrup, H. P. and Larsen, E. (2003) 'Lateglacial landform associations at Jaeren (SW Norway) and their glacio-dynamic implications.' *Boreas*, **32**, 462 – 475.

Reynolds, J. M. (2011) *An introduction to applied and environmental geophysics*. Wiley: Oxford

Rose, J. (1989) 'Glacier stress patterns and sediment transfer associated with the formation of superimposed flutes.' *Sedimentary Geology*, **62** (2), 151 – 176.

Sahlin, E. A. and Glasser, N. F. (2008) 'A geomorphological map of Cadair Idris, Wales.' *Journal of Maps*, **4** (1), 299 – 314.

Schoof, C. (2007) 'Pressure-dependent viscosity and interfacial instability in coupled ice-sediment flow.' *Journal of Fluid Mechanics*, **570**, 227 – 252.

Shaw, J. and Gilbert, R. (1990) 'Evidence for large-scale subglacial meltwater flood events in southern Ontario and northern New York state.' *Geology*, **18**, 1169 – 1172.

Shaw, J. and Kvill, D. (1984) 'A glaciofluvial origin for drumlins of the Livingstone Lake area, Saskatchewan.' *Canadian Journal of Earth Science*, **21**, 1442 – 1459.

Shaw, J., Kvill, D. and Rains, B. (1989) 'Drumlins and catastrophic subglacial floods.' *Sedimentary Geology*, **62**, 177 – 202.

Shepherd, A., Ivins, E. R., Geruo, O., Barletta, V. R., Bentley, M. J., Bettadpur, S. *et al*. (2012) 'A reconciled estimate of ice-sheet mass balance.' *Science*, **338** (6111), 1183 – 1189.

Smalley, I. J. and Unwin, D. J. (1968) 'The formation and shape of drumlins and their distribution and orientation in drumlin fields.' *Journal of Glaciology*, **7**, 377 – 390.

Smith, A. M., Murray, T., Nicholls, K. W., Makinson, K., Adalgeirsdottir, G., Behar, A. E. and Vaughan, D. G (2007) 'Rapid erosion, drumlin formation, and changing hydrology beneath an Antarctic ice stream.' *Geology*, **35** (2), 127 – 130.

Spagnolo, M., Clark, C. D. and Hughes, A. L. C. (2012) 'Drumlin relief.' *Geology*, **153**, 179 – 191.

Spagnolo, M., Clark, C. D., Hughes, A. L. C. and Dunlop, P. (2011) 'The topography of drumlins: assessing their long profile shape.' *Earth Surface Processes and Landforms*, **36**, 790 – 804.

Spagnolo, M., Clark, C. D., Hughes, A. L. C., Dunlop, P. and Stokes, C. R. (2010) 'The planar shape of drumlins.' *Sedimentary Geology*, **232**, 119 – 129.

Spagnolo, M., King, E. C., Ashmore, D. W., Rea, B. R., Ely, J. C., and Clark, C. D. (2014) 'Looking through drumlins: testing the application of ground-penetrating radar.' *Journal of Glaciology,* **60** (224), 1126 – 1134.

Stokes, C. R. and Clark, C. D. (2001) 'Palaeo-ice streams.' *Quaternary Science Reviews*, **20** (13), 1437 – 1457.

Stokes, C. R. and Clark, C. D. (2002) 'Are long subglacial bedforms indicative of fast ice flow?' *Boreas,* **31** (3), 239 – 249.

Stokes, C. R., Spagnolo, M. and Clark, C. D. (2011) 'The composition and internal structure of drumlins: Complexity, commonality, and implications for a unifying theory of their formation.' *Earth-Science Reviews,* **107** (3), 398 – 422.

Sugden, D. E. and John, B. S. (1976) *Glaciers and Landscape: A geomorphological approach*. Edward Arnold, London.

Thackeray, S. Maberly, S. and Winfield, I. (2010) 'The ecology of Bassenthwaite Lake (English Lake District).' *Freshwater Forum*, **25**.

Thomas, G. S. and Connell, R. J. (1985) 'Iceberg drop, dump, and grounding structures from Pleistocene glaciolacustrine sediments, Scotland.' *Journal of Sedimentary Research*, **55** (2).

Truffer, M., Echelmeyer, K. A. and Harrison, W. D. (2001) 'Implications of till deformation on glacier dynamics.' *Journal of Glaciology,* **47** (156), 123 – 134.

Turner, A. J., Woodward, J., Dunning, S. A., Shine, A. J., Stokes, C. R. and O'Cofaigh, C. (2012) 'Geophysical surveys of the sediments of Loch Ness, Scotland: Implications for the deglaciation of the Moray Firth Ice Stream, British-Irish ice sheet.' *Journal of Quaternary Science*, **27**, 221 – 232.

Van Landeghem, K. J., Wheeler, A. J. and Mitchell, N. C. (2009) 'Seafloor evidence for palaeo-ice streaming and calving of the grounded Irish Sea Ice Stream: implications for the interpretation of its final deglaciation phase.' *Boreas*, **38** (1), 119 – 131.

Waldmann, N., Ariztegui, D., Anselmetti, F. S., Coronato, A. and Austin, J. A. (2010) 'Geophysical evidence of multiple glacier advances in Lago Fagnano (54^{o}S), southernmost Patagonia.' *Quaternary Science Reviews*, **29** (9), 1188 – 1200.

Warren, C. and Aniya, M. (1999) 'The calving glaciers of southern South America.' *Global and Planetary Change*, **22** (1), 59 – 77